

A Geological Basis for the Exploration of the Planets

**CASE FILE
COPY**



NATIONAL AERONAUTICS
AND SPACE ADMINISTRATION

A Geological Basis for the Exploration of the Planets

Edited by

RONALD GREELEY
*University of Santa Clara
NASA Ames Research Center*

and

MICHAEL H. CARR
U.S. Geological Survey



Scientific and Technical Information Office
1976
NATIONAL AERONAUTICS AND SPACE ADMINISTRATION
Washington, D.C.

For sale by the National Technical Information Service
Springfield, Virginia 22161
Price - \$5.25

PREFACE

This document was first prepared in the late 1960's at the request of the NASA Planetology Program Office and was published under the title, "A Strategy for the Geologic Exploration of the Planets," edited by M. H. Carr (1970). The objectives were to define the goals of planetary geology and to set forth the methods of meeting those goals. The report has been used by the planetary geologist as a reminder of the relationship of his specialized pursuits to the long-term objectives of planetology and by non-geologists as a means of understanding the rationale for and the limitations of planetary geologic data. The primary intended use for such a "strategy," however, was its use as a planning document to help define the kinds of missions that should be developed and the priorities of those missions in the geologic context and to aid in identifying the supporting research requirements.

The objectives of the original strategy have not changed, nor have the long-term objectives of planetary geology and their relation to space science. The general approach in geology has evolved over several hundred years and the approach appears to work as well on other planets as it does on Earth. However, the original exploration strategy document was written essentially in pre-Apollo times; since then there have been six manned and three unmanned Moon landings, two flybys and one orbiter of Mars, two flybys of Jupiter (and "fly-throughs" of the asteroid belt), a flyby of Venus and Venus landers, and a flyby (actually three flybys, or encounters, by the same spacecraft) of Mercury. Additionally, a great deal of Earth-based data has been acquired, such as radar images for parts of the surface of Venus. Instrument development, especially miniaturization and on-board data processing, has increased the number of experiments that can be performed during each mission. In light of these tremendous advances in space science, there is a need to re-examine the specific objectives of the field and the methods for achieving those objectives.

This report considers the geologic aspects of solar-system studies by first showing how geologic data are related to space science in general and, second, by discussing the approach generally used in *planetary geology*. Planetary geology is the study of the origin, evolution, and distribution of matter condensed in the form of planets, satellites, comets, and asteroids. It is a multidisciplinary effort involving, among others, investigators with backgrounds in geology, chemistry, physics, and astronomy. More than any other endeavor, exploration of the solar system requires close communication among all disciplines. The term "geology" is used here in its broadest sense and is considered to mean the study of the solid parts of the planets. Geophysics, geochemistry, geodesy, cartography, and other disciplines concerned with the solid planets are all included in the general term.

Planetary atmospheres are not covered, except in relation to their evolution by the release of volatiles through geologic processes and in relation to surface processes such as dust storms and fluvial processes. Consequently, much of the discussion is limited to the terrestrial planets and to the solid satellites of the outer planets. Jupiter and Saturn, in particular, are difficult to consider in geologic terms because they probably have no solid surfaces in the usual sense; however, their satellites are considered to be of prime importance.

This report is not concerned with any one mission, or series of missions, nor does it outline a sequence of planet exploration. Rather, it is restricted to the kinds of experiments and observations made through unmanned missions. Because planetary mission planning is complicated by the necessary interactions of scientific objectives, fiscal constraints, and the limited opportunities for exploration caused by the planets' orbital peculiarities (e.g., the optimum alignment of the outer planets for missions from Earth occurs only once in each 180 yr), it is imperative that future mission planners have inputs from a broad spectrum of scientists.

The sections that follow were written by contributors (Appendix I) who typically are experts in their particular discipline. In many cases, the contributor has had on-board spacecraft experiments and is thus in a position to know the advantages, as well as the limitations, of the data. While we recognize the impossibility of obtaining complete agreement by the geologic community on the methods of planetary geology, we do feel that this document represents a consensus within the framework presented.

The report has been written partly within the framework of the recommendations of the Woods Hole Conference (National Academy of Sciences, 1966) and the MacDonald Committee on Space Research (National Academy of Sciences, 1968). We have relied on several other reports for guidance, particularly those of Adams *et al.* (1967a,b); Mutch and Soffen (unpublished report); Wetherill *et al.* (1973); and Goody *et al.* (1974). This report differs from most of the others, however, in its emphasis on geologic problems.

REFERENCES CITED

Adams, J. B., Conel, J. E., Dunne, J. A., Fanale, F., Holstrom, G. B., and Loomis, A. A., 1967a. A strategy for scientific exploration of the terrestrial planets. *Rev. Geophys.* 7:623-661.

Adams, J. B., Conel, J. E., Dunne, J. A., Fanale, F., Holstrom, G. B., and Loomis, A. A., 1967b. Scientific strategy, objectives and experiments for Mariner missions from 1970 through 1973. Calif. Inst. Tech. Jet Propulsion Lab. Doc. 900-94 (internal document).

Anon., 1966. Space research — directions for the future. Natl. Acad. Sci.-Natl. Res. Council Pub. 1403.

Anon., 1968. Planetary exploration 1969-1975. Natl. Acad. Sci., Washington, D.C.

Carr, M. H., editor, 1970. A strategy for the geologic exploration of the planets. U.S. Geol. Survey Circular 640.

Goody, R. M., Chairman, 1974. Opportunities and choices in space sciences, 1974. Space Science Board, National Research Council.

Wetherill, G., chairman, 1973. Mars: A strategy for exploration. A report of a study by the Mars Science Advisory Committee to the Planetary Programs Office, NASA.

ACKNOWLEDGMENTS

The editors have drawn on the abilities and resources of many individuals in the preparation of this report. We thank first and foremost the contributing authors who, despite certain time constraints, generally supplied their manuscripts on schedule. We also acknowledge the efforts and careful considerations of the many reviewers, particularly, A. Albee, W. Cameron, G. Fielder, D. Gault, C. Hodges, W. Kaula, J. O'Keefe, R. Phillips, C. Sagan, P. Schultz, W. Sjogren, A. Turkevich, and F. Whipple.

The work was supported by the Planetary Geology Program, NASA, through Contract W-13,576 (M.H.C.) and Grant NGR-05-017-037 (R.G.), and by the Space Sciences Division, Ames Research Center, NASA.

Page intentionally left blank

Page intentionally left blank

TABLE OF CONTENTS

| Chapter | Page |
|--|------|
| PREFACE | iii |
| ACKNOWLEDGMENTS | v |
| | |
| I. INTRODUCTION – <i>R. Greeley and M. H. Carr</i> | 1 |
| RELEVANCE OF PLANETARY GEOLOGY | 1 |
| Origin and Evolution of the Solar System | 1 |
| Origin and Evolution of Life | 6 |
| Man's Terrestrial Environment | 6 |
| ASTEROIDS AND COMETS – <i>C. R. Chapman</i> | 7 |
| PLANETARY GEOLOGY APPROACH | 9 |
| Present Geologic State | 10 |
| Geologic Evolution of Planets | 10 |
| Comparative Planetology | 10 |
| REFERENCES CITED | 11 |
| | |
| II. STRATIGRAPHY AND STRUCTURAL GEOLOGY – <i>M. H. Carr, D. E. Wilhelms, R. Greeley, and J. E. Guest</i> | 13 |
| PRINCIPLES | 13 |
| Introduction | 13 |
| Heterogeneities of Planetary Surfaces – A Stratigraphic Problem | 14 |
| Geologic Framework | 15 |
| Methodology | 15 |
| DATA REQUIREMENTS | 20 |
| Image Parameters | 20 |
| Spatial resolution | 20 |
| Illumination | 20 |
| Color | 21 |
| Stereoscopic coverage | 22 |
| Polarimetry | 22 |

| | | |
|--|-----|-----|
| Types of Missions | .22 | |
| Flyby missions | .22 | |
| Orbital missions | .23 | |
| Landing missions (<i>E. Morris</i>) | .23 | |
| Radar images (<i>G. Schaber</i>) | .25 | |
| IMAGING STRATEGY— <i>B. Murray</i> | .26 | |
| Areal Coverage | .27 | |
| Full Disk Images | .29 | |
| Contiguous Coverage | .30 | |
| Nested High Resolution Frames | .30 | |
| SUMMARY | .30 | |
| REFERENCES CITED | .31 | |
| III. GEOCHEMISTRY — <i>M. H. Carr and P. Toulmin III</i> | | .33 |
| ELEMENTAL COMPOSITION | .35 | |
| Sampling | .35 | |
| Accuracy and Precision | .36 | |
| Analytical Techniques | .37 | |
| ISOTOPIC COMPOSITION — <i>R. Zartman</i> | .38 | |
| MINERAL COMPOSITION | .39 | |
| ORGANIC CHEMISTRY — <i>K. A. Kvenvolden</i> | .40 | |
| PLANETARY VOLATILES — <i>F. P. Fanale</i> | .41 | |
| WATER — <i>D. M. Anderson</i> | .45 | |
| Summary | .49 | |
| ORBITAL MEASUREMENTS — <i>M. H. Carr</i> | .49 | |
| Gamma Ray Spectroscopy (<i>A. Metzger</i>) | .50 | |
| X-Ray Spectroscopy (<i>A. Metzger</i>) | .51 | |
| Far-Ultraviolet Radiometry (<i>M. A. Steggart</i>) | .52 | |
| Multiband Photometry in the Visible and Near-Infrared (<i>J. Adams, T. McCord, and R. Huguenin</i>) | .52 | |
| Far-Infrared Spectroscopy | .54 | |
| Infrared Radiometry (<i>H. Kieffer</i>) | .55 | |
| Microwave Radiometry | .56 | |
| SUMMARY | .57 | |
| REFERENCES CITED | .58 | |

| | |
|--|-----|
| IV. GEOPHYSICS – <i>M. H. Carr and P. Cassen</i> | 63 |
| GRAVITY – <i>S. Saunders</i> | 64 |
| MAGNETOMETRY | 66 |
| SEISMOLOGY | 67 |
| HEAT FLOW – <i>M. Langseth</i> | 69 |
| RADAR STUDIES OF SURFACE PROPERTIES – <i>L. Tyler, T. Howard, A. W. England, and J. Cuzzi</i> | 71 |
| REFERENCES CITED | 73 |
| V. GEODESY AND CARTOGRAPHY – <i>R. Batson and D. Arthur</i> | 75 |
| GEODESY | 75 |
| CARTOGRAPHY | 76 |
| Data Acquisition | 76 |
| Computation of Controls | 76 |
| Map Compilation | 78 |
| DATA ACQUISITION SYSTEMS | 78 |
| MISSION DESIGN | 81 |
| Geodesy | 81 |
| Map Control and Cartography | 81 |
| DATA REDUCTION | 82 |
| SUMMARY | 82 |
| REFERENCES CITED | 83 |
| VI. EARTH-BASED STUDIES – <i>S. Dwornik</i> | 85 |
| FIELD STUDIES – <i>D. U. Wise and G. E. McGill</i> | 88 |
| LABORATORY EXPERIMENTATION – <i>A. Howard</i> | 91 |
| EXPERIMENTAL PETROLOGY – <i>M. C. Gilbert</i> | 94 |
| Introduction | 94 |
| Relevance of Experimental Petrology to Planetology | 94 |
| METEORITES – <i>T. Bunch</i> | 97 |
| Summary | 100 |
| GROUND-BASED TELESCOPIC OBSERVATIONS – <i>E. Helin, D. Matson, and T. Johnson</i> | 100 |

| | |
|--|-----|
| TERRESTRIAL RADAR OBSERVATIONS – <i>G. Schaber</i> | 102 |
| THEORETICAL STUDIES | 104 |
| REFERENCES CITED | 104 |
| | |
| APPENDIX I. CONTRIBUTING AUTHORS | 107 |

CHAPTER I. INTRODUCTION

R. Greeley and M. H. Carr

RELEVANCE OF PLANETARY GEOLOGY

Exploration has always been held in fascination; geographic exploration of unknown territory engenders excitement in even the most casual observer. Exploration of unknown areas today requires the efforts of many specialists, each contributing a part to the whole and each sharing the rewards. As the third decade of the space age is approached, the exploration of the solar system is becoming a reality (Tables 1.1-1.3; Fig. 1.1).

This report considers the geologic aspects of solar-system exploration by first showing how geologic data are related to space science in general and, second, by discussing the approach generally used in *planetary geology*. Planetary geology is the study of the origin, evolution, and distribution of matter condensed in the form of planets, satellites, comets, and asteroids. It is a multidisciplinary effort involving, among others, investigators with backgrounds in geology, chemistry, physics, and astronomy. More than any other endeavor, exploration of the solar system requires close communication between all disciplines. The term "geology" is used here in its broadest sense and is considered to mean the study of the solid parts of the planets. Geophysics, geochemistry, geodesy, cartography, and other disciplines concerned with the solid planets are all included in the general term.

Planetary atmospheres are not covered, except in relation to their evolution by the

release of volatiles through geologic processes and in relation to surface processes such as dust storms and fluvial activity. Consequently, much of the discussion is limited to the terrestrial planets, comets, and asteroids, and to the solid satellites of the outer planets. Jupiter and Saturn, in particular, are difficult to consider in geologic terms because they probably have no solid surfaces in the usual sense; however, their satellites are considered to be of prime importance.

The National Academy of Sciences (1966) placed three goals before the scientific community: (1) to determine the origin and evolution of the solar system, (2) to determine the origin and evolution of life, and (3) to clarify the nature of the processes that shape man's terrestrial environment. These goals form part of NASA's basic charter of space exploration, and planetary geology figures prominently in all three of these ambitious goals.

Origin and Evolution of the Solar System

There are two ways of looking at the origin and evolution of the solar system: one way is to attempt to model the solar system by following a series of stages in its development to the present condition; the other is to examine the present state of the solar system and to interpret its history by moving backward in time. Although both methods are valuable, it is the latter one with which the geologist is concerned.

TABLE 1.1. SUCCESSFUL PLANETARY MISSIONS LAUNCHED BY THE UNITED STATES AND THE SOVIET UNION (AFTER SAGAN 1975)

| NAME OF SPACECRAFT | DATE OF LAUNCH | DESTINATION | DATE OF ENCOUNTER | NEAREST APPROACH (KILOMETERS) | STATUS OF MISSION |
|--------------------|--------------------|----------------|-------------------------|-------------------------------|---|
| MARINER 2 | 8/26/62 | VENUS | 12/14/62 | 35,000 | First flyby of another planet; found high temperature (400 degrees Celsius) arises from surface not atmosphere; no evidence of magnetic field. |
| MARINER 4 | 11/28/64 | MARS | 7/14/65 | 10,000 | First flyby of Mars; returned 22 television pictures of Martian surface, other data. |
| VENERA 4 | 6/12/67 | VENUS | 10/18/67 | LANDED | First on-site measurements of temperature, pressure and composition of Venusian atmosphere; probe transmitted data during 94-minute parachute descent. |
| MARINER 5 | 6/14/67 | VENUS | 10/19/67 | 4,000 | Measured structure of upper atmosphere of Venus during flyby. |
| VENERA 5 | 1/5/69 | VENUS | 5/16/69 | LANDED | Probes transmitted data on pressure, temperature and composition of atmosphere during parachute descent; missions similar to that of <i>Venera 4</i> . First successful landing on another planet. |
| VENERA 6 | 1/10/69 | VENUS | 5/17/69 | LANDED | |
| MARINER 6 | 2/25/69 | MARS | 7/31/69 | 3,390 | Flyby obtained infrared and ultraviolet spectra of atmosphere; returned 76 pictures of surface, other data. |
| MARINER 7 | 3/27/69 | MARS | 8/5/69 | 3,500 | Mission identical with that of <i>Mariner 6</i> ; returned 126 pictures of surface, 33 of south-polar region. |
| VENERA 7 | 8/17/70 | VENUS | 12/15/70 | LANDED | Mission similar to those of <i>Venera 4</i> , <i>Venera 5</i> and <i>Venera 6</i> . |
| MARS 3 | 5/28/71 | MARS | 12/2/71 | LANDED | Orbiter achieved Mars orbit and returned data; descent module soft-landed and transmitted 20 seconds of featureless television data before failing. |
| MARINER 9 | 5/30/71 | MARS | 11/13/71 | 1,395 | First spacecraft to go into orbit around another planet; returned 7,329 pictures of surface, atmosphere, clouds and satellites, together with other data. |
| PIONEER 10 | 3/3/72 | JUPITER | 12/4/73 | 131,400 | Successfully traversed asteroid belt; investigated interplanetary medium, Jovian magnetosphere and atmosphere; returned more than 300 pictures of Jovian clouds and satellites; first spacecraft to use gravity-assisted trajectory; first man-made object to escape solar system |
| VENERA 8 | 3/26/72 | VENUS | 7/22/72 | LANDED | Survived Venusian surface conditions for 50 minutes; determined radioactive content of surface; on entry measured winds and sunlight penetrating clouds. |
| PIONEER 11 | 4/6/73 | JUPITER SATURN | 12/3/74 (J) 9/79 (S) | 46,400 (J) | Second Jupiter flyby; now en route to Saturn, then to leave solar system. |
| MARS 4 MARS 5 | 7/21/73 7/25/73 | MARS | 1/74 1/74 | ? | Went into orbit around Mars; returned photographs of surface and other data. |
| MARS 6 | 8/5/73 | MARS | 2/74 | LANDED | Descent module failed at touchdown; entry data suggest high argon content of atmosphere. |
| MARINER 10 | 11/3/73 | VENUS MERCURY | 2/5/74 3/29/74 (M) | 5,800 (V) 700 (M) | First probe of Mercury; returned more than 8,000 pictures and other data from Venus and Mercury; re-encountered Mercury 9/21/74 and 3/16/75. |
| VENERA 9 | 6/8/75 | VENUS | 10/22/75 | LANDED | Survived 53 minutes; returned panoramic picture of surface; measured surface pressure of 90 atm; and temperature of 485° C. |
| VENERA 10 | 6/14/75 | VENUS | 10/25/75 | LANDED | Survived 65 minutes; returned panoramic picture of surface; measured surface pressure of 92 atm; and temperature of 465° C. |
| VIKING 1 | 8/75 | MARS | 6/76 | TO LAND | Orbiter to study atmosphere and photograph surface; lander to study atmosphere at surface, investigate surface geology and chemistry and test soil for signs of extraterrestrial life. |
| VIKING 2 | 9/75 | MARS | 8/76 | TO LAND | |

| | | | | | |
|------------|------|----------------|----------------------|---------|---|
| MARINER 11 | 8/77 | JUPITER SATURN | 1979 (J) 1981 (S) | ? | To conduct comparative studies of two outer planets and their 23 satellites; to investigate nature of Saturn's rings; to measure interplanetary medium out to Saturn's orbit; 20,000 photographs planned. |
| MARINER 12 | 9/77 | JUPITER SATURN | 1979 (J) 1981 (S) | ? | |
| PIONEER 12 | 5/78 | VENUS | 12/78 | TO LAND | Orbiter to study interaction of atmosphere with solar wind over one 243-day period; "bus" to drop three small probes toward surface, then relay data to earth as they enter atmosphere. |
| PIONEER 13 | 8/78 | VENUS | 12/78 | TO LAND | |

TABLE 1.2. CHARACTERISTICS OF THE MAJOR PLANETS (AFTER SAGAN 1975)

| | MERCURY | VENUS | EARTH | MARS | JUPITER | SATURN | URANUS | NEPTUNE | PLUTO |
|--|-------------------------------|--------------------|------------------|---------------------------|------------------|---------------------------|---------------------------|---------------------------|---------------|
| MEAN DISTANCE FROM SUN (MILLIONS OF KILOMETERS) | 57.9 | 108.2 | 149.6 | 227.9 | 778.3 | 1,427 | 2,869.6 | 4,496.6 | 5,900 |
| PERIOD OF REVOLUTION | 88 DAYS | 224.7 DAYS | 365.26 DAYS | 687 DAYS | 11.86 YEARS | 29.46 YEARS | 84.01 YEARS | 164.8 YEARS | 247.7 YEARS |
| INCLINATION OF AXIS | < 28° | 3° | 23° 27' | 23° 59' | 3° 05' | 26° 44' | 82° 5' | 28° 48' | ? |
| EQUATORIAL DIAMETER (KILOMETERS) | 4,880 | 12,104 | 12,756 | 6,787 | 142,800 | 120,000 | 51,800 | 49,500 | 6,000 (?) |
| MASS (EARTH = 1) | .055 | .815 | 1 | .108 | 317.9 | 95.2 | 14.6 | 17.2 | .1 (?) |
| VOLUME (EARTH = 1) | .06 | .88 | 1 | .15 | 1,316 | 755 | 67 | 57 | .1 (?) |
| DENSITY (WATER = 1) | 5.4 | 5.2 | 5.5 | 3.9 | 1.3 | .7 | 1.2 | 1.7 | ? |
| ATMOSPHERE (MAIN COMPONENTS) | NONE | CARBON DIOXIDE | NITROGEN, OXYGEN | CARBON DIOXIDE, ARGON (?) | HYDROGEN, HELIUM | HYDROGEN, HELIUM, METHANE | HYDROGEN, HELIUM, METHANE | HYDROGEN, HELIUM, METHANE | NONE DETECTED |
| MEAN TEMPERATURE AT VISIBLE SURFACE (DEGREES CELSIUS) S = SOLID, C = CLOUDS | 350 (S) DAY -170 (S) NIGHT | -33 (C) 480 (S) | 22 (S) | -23 (S) | -150 (C) | -180 (C) | -210 (C) | -220 (C) | -230 (?) |
| ATMOSPHERIC PRESSURE AT SURFACE (MILLIBARS) | 10 ⁹ | 90,000 | 1,000 | 6 | ? | ? | ? | ? | ? |
| SURFACE GRAVITY (EARTH = 1) | .37 | .88 | 1 | .38 | 2.64 | 1.15 | 1.17 | 1.18 | ? |
| KNOWN SATELLITES | 0 | 0 | 1 | 2 | 13 | 10 | 5 | 2 | 0 |

TABLE 1.3. CHARACTERISTICS OF THE SATELLITES (AFTER ALLEN 1964)

| Planet | Satellites | Distance from planet, thousands of km | Sidereal period,* days | Synodic period,** d h m | Orbit inclination,*** deg | Orbit eccentricity | Radius, km | Mass 10^{24} g |
|---------|-------------|---------------------------------------|------------------------|-------------------------|---------------------------|--------------------|--------------------|------------------|
| Earth | Moon | 384 | 27.321 661 | 29 12 44 | 23 | 0.054 9 | 1 738 | 73.5 |
| Mars | 1 Phobos | 9 | 0.318 910 | 7 39 | 1 | 0.021 | 14 by 10 7 by 6 | 73 48 |
| | 2 Deimos | 23 | 1.262 441 | 1 06 21 | 2 | 0.002 8 | | |
| Jupiter | 1 Io | 422 | 1.769 138 | 1 18 29 | 0 | small | 1 810 | 73 |
| | 2 Europa | 671 | 3.551 181 | 3 13 18 | 1 | and | 1 480 | 48 |
| | 3 Ganymede | 1 070 | 7.154 553 | 7 04 00 | 0 | vari- | 2 600 | 154 |
| | 4 Callisto | 1 883 | 16.689 018 | 16 18 05 | 0 | able | 2 360 | 95 |
| | 5 Amalthea | 181 | 0.418 178 | 11 57 | 0 | 0.003 | 80 | |
| | 6 Himalia | 11 476 | 250.566 | 265 23 | 28 | 0.158 | 50 | |
| | 7 Elara | 11 737 | 259.65 | 276 05 | 26 | 0.206 | 12 | |
| | 8 Pasiphae | 23 500 | 737 | 634 | 33 R | 0.40 | 10 | |
| | 9 Sinope | 23 600 | 758 | 645 | 24 R | 0.27 | 9 | |
| | 10 Lysithea | 11 700 | 259.22 | 275 17 | 29 | 0.135 | 8 | |
| | 11 Carme | 22 600 | 692 | 297 | 17 R | 0.207 | 9 | |
| | 12 Ananke | 21 200 | 630 | 551 | 33 R | 0.16 | 8 | |
| | 13 Leda | 12 400 | 282 | | | | | |
| | 14 | | | | | | | |
| Saturn | 1 Mimas | 186 | 0.942 422 | 22 37 | 2 | 0.020 1 | 270 | 0.04 |
| | 2 Enceladus | 238 | 1.370 218 | 1 08 53 | 0 | 0.004 4 | 300 | 0.08 |
| | 3 Tethys | 295 | 1.887 802 | 1 21 19 | 1 | 0.0 | 500 | 0.64 |
| | 4 Dione | 377 | 2.736 915 | 2 17 42 | 0 | 0.002 2 | 480 | 1.1 |
| | 5 Rhea | 527 | 4.417 503 | 4 12 28 | 0 | 0.001 0 | 650 | 2.3 |
| | 6 Titan | 1 222 | 15.945 449 | 15 23 15 | 0 | 0.029 0 | 2 440 | 137 |
| | 7 Hyperion | 1 481 | 21.276 657 | 27 07 39 | 1 | 0.104 | 220 | 0.1 |
| | 8 Iapetus | 3 560 | 79.330 84 | 79 22 05 | 15 | 0.028 3 | 550 | 1.1 |
| | 9 Phoebe | 12 950 | 550.33 | 523 13 36 | 30 R | 0.163 3 | 120 | |
| | 10 Janus | 159 | 0.749 0 | 17 59 | 0 | 0.0 | 150 | |
| | 11 Rings | 73 to 137 | | | | | 10 thickness | <28 |
| Uranus | 1 Ariel | 192 | 2.520 38 | 2 12 30 | 0 | 0.003 | 350 | 1.3 |
| | 2 Umbriel | 267 | 4.144 18 | 4 03 28 | 0 | 0.004 | 250 | 0.5 |
| | 3 Titania | 438 | 8.705 88 | 8 17 00 | 0 | 0.002 4 | 500 | 4.3 |
| | 4 Oberon | 586 | 13.463 26 | 13 11 16 | 0 | 0.000 7 | 450 | 2.6 |
| | 5 Miranda | 128 | 1.414 | | | <0.01 | 120 | 0.1 |
| Neptune | 1 Triton | 353 | 5.876 83 | 5 21 03 | 20 R | 0.0 | 1900 | 140 |
| | 2 Nereid | 5 600 | 359.88 | 362 01 | 28 | 0.76 | 120 | 0.03 |

**Sidereal period* is the time between two successive returns of the satellite to the same point in the heavens as seen from the planet.

***Synodic period* is the time between two successive returns of the satellite to the same illumination (phase) as seen from the planet.

***"R" indicates retrograde motion.

Adams *et al.* (1967a,b) examined the problem of the origin and evolution of the solar system in the geological context and reduced it to these questions:

1. Are the individual terrestrial planets and satellites chemically uniform or non-uniform?
2. Did the final accretion result in the present array of planets and satellites, or in an array that was subsequently altered?
3. Was the cloud chemically homogeneous at the time of final accretion?
4. What was the state of the Sun-cloud system when it first appeared as a recognizable unit?
5. Were there large-scale elemental and isotopic nonuniformities in the contracted nebula?

These questions concern the determination of the stage (or stages) in solar-system evolution when chemical fractionation took place, so answers must depend ultimately on chemical measurements as well as mineralogy and petrology. Many of the necessary measurements will be made on planets and their satellites, and their interpretation will confront us with the problem of chemical inhomogeneity. Clearly, we cannot assume *a priori* that a planet is chemically homogeneous and that a given analysis is representative of the whole planet; this is not true of the Earth and Moon but appears to be true for Mars and Mercury (see section on "Heterogeneities of Planetary Surfaces," Ch. II). Neither can we simply assume random variability and sample a planet statistically. The only accessible part of a planet — its near-surface — may be totally



Lick Observatory Photo L-9

Figure 1.1. *Lunar exploration via successful manned and unmanned landings. Roman numerals in circles are unmanned U.S. Surveyor spacecraft, numbers in circles are U.S. manned Apollo landing sites, triangles are Soviet unmanned Lunokhod sites, and squares are Soviet unmanned Luna sites. Not shown are "hard" landings such as the U.S. Ranger impact areas (modified from Lunar Science Institute map).*

nonrepresentative of the planet as a whole. We must have a general understanding of the planet, and this can be achieved effectively only by studying the possible processes that may have caused – and may yet be causing – redistribution

of materials within the planet. Only if the processes are understood can valid general conclusions be made regarding the significance of specific measurements. The sampling or type of measurement made on a planet therefore should

be guided by geologic priorities, based on this need to understand geologic processes.

Planetary geophysics is concerned largely with documenting and interpreting internal heterogeneities and conditions. Chemical determinations of materials from the interiors obviously cannot be made directly; nevertheless, inferences are possible from measurements of physical properties. The value of any particular physical parameter is not important in itself since it depends on local conditions; its importance lies in the fact that it places limits on chemical composition and internal conditions and provides data on internal processes. Stratigraphy is concerned directly with documentation of crustal heterogeneity. Even though the surface rocks constitute a minute part of the total mass of the planet, it is the part on which a great many measurements are made. For this reason, an understanding of surface materials is crucial. We must know how variegated the surface rocks are and how they were formed. The mode of formation is particularly important, for different processes vary in the extent to which they cause chemical change. Consequently, to understand the broader implications of surface analyses, we should know the distribution and mode of formation of the rocks analyzed, as well as the degree to which they typify all the materials of the planet.

Origin and Evolution of Life

The geology of planetary surfaces bears directly on the origin of life. This report has been written largely on the assumption that, in our solar system, life will not be found on other planets. But if life is found elsewhere, then the significance of geology will be enhanced and this strategy will need drastic revision. With life on a planet, such factors as the mode of rock formation, the physical conditions of deposition, and the relative age will take on new significance. Not only will these factors reflect past conditions and events, but they will reveal the evolutionary path of life on the planet and the

conditions under which life thrived. Paleontology, which has been ignored in this strategy, would become a study of paramount importance, and different types of missions would be required to meet its needs.

Man's Terrestrial Environment

One important result of planetary exploration will be increased knowledge of the Earth. Many fundamental geologic problems could be solved by detailed comparison of the Earth with other planetary bodies. The relative effects of size, original composition, and the presence of an atmosphere and hydrosphere on the evolution of the Earth are of particular geologic importance; comparison of the Earth with other bodies will allow these effects to be assessed.

Probably the most important result of planetary exploration from the point of view of man's terrestrial environment will be an improved understanding of the early history and deep interior of the Earth. Because evidence of events and processes that took place early in Earth's history rarely survives, very little is known of the first two aeons of the Earth's history. Features on the Earth's surface are subject to relatively rapid destruction and modification, largely as a result of the erosive action of water, but also as the result of tectonic and volcanic activity. We now know that the early surface history of both the Moon and Mercury has been preserved, partly because they have been relatively inactive internally through much of their later existence; moreover, because neither has an atmosphere or free water, erosion has been gradual. The situation on Mars is similar, although the surface of Mars has been locally modified by both internal activity and surface processes. Because all these planets appear to have had a similar early history, the analogy between them and the early history of the Earth is strengthened. Indeed, the only way to arrive at an understanding of the Earth's early history may be by analogy with what is learned of these more primitive planets.

ASTEROIDS AND COMETS

C. R. Chapman

Asteroids, comets, and other small bodies have importance for solar system studies proportionately far greater than their cumulative mass. Whereas the planets and larger satellites have experienced geochemical differentiation, volcanism, weathering, geomorphological modification, and accumulation of vast impact regoliths that all serve to modify and bury the rocks originally present, comets and asteroids are more likely to have retained solar nebular condensates and other early materials in relatively unmodified form. Furthermore, these materials are likely to be more readily accessible for analysis because the low surface gravity means that deep regoliths cannot accumulate so that fresh materials are continually being exposed on the surfaces of these bodies.

Current models consider the planets to have accreted from smaller "planetesimals" of which the asteroids and comets may be leftover examples. There are increasing indications that many early planetesimals that formed in the vicinity of the major planets (especially Jupiter) were scattered into other parts of the solar system during and subsequent to planetary accretion. Therefore, the surfaces of other planets (especially Mars) may consist of material that condensed originally in a different part of the solar nebula.

The volatile-rich comets are presumably the most pristine examples of early planetesimals. They arrive in the inner solar system from time to time from cold storage in the "Oort comet cloud," a cloud postulated to encircle the outer fringes of the solar system. In that environment, they have been protected both from significant bombardment as well as from proximity to the Sun. As they approach the Sun for the first time, the outer layers of "new" comets begin to vaporize, forming the characteristic comae and tails composed of neutral and ionized gases and dust. From astronomical observations of these released components, inferences are made about the nature and composition of cometary nuclei (probably they are assemblages of ices and rocks).

Geological interest in comets bears on the nature of the nucleus, which is a small object and cannot be examined closely until there is a spacecraft rendezvous with a comet. It is not yet known which meteorites, if any, have come from comets; Type I carbonaceous chondrites are the most likely candidates. The chief advantage of studying comets is that they are so obviously composed of unaltered, "primitive" material. But the same orbital processes that have preserved these materials for 4.5 aeons and then brought them back to our vicinity also have destroyed any clear record of where the comets first formed. So we do not know if the materials contained in comets first condensed near Jupiter, near Neptune, or elsewhere in the solar system.

The best evidence suggests that the asteroids, on the other hand, are located in orbits rather similar to those in which they were formed — mostly between Mars and Jupiter. Thus the geochemical sampling of asteroids at different semi-major axes can provide information on solar nebular condensates that accreted between the Earth and Jupiter. But the asteroids have paid a price for their proximity to the inner solar system. At least a few of them, notably Vesta (McCord *et al.* 1970) and probably the 10 percent of the larger ones classified as stony-iron objects (McCord and Gaffey 1974; Chapman 1974), have undergone geochemical differentiation. Moreover, the latest models for the collisional evolution of asteroids suggest that nearly all of the asteroids observed today are collisional fragments of precursor planetesimals (Chapman and Davis 1975). Still, most asteroids appear to be mineralogically similar to the relatively unaltered carbonaceous chondrites and a few others seem similar to the ordinary chondrites and other meteorite types (Chapman 1975).

In addition, the collisions have probably not altered the asteroidal rocks appreciably and, in fact, serve to place some fragments into Earth-crossing orbits to fall as meteorites. The orbital processes that transfer asteroidal fragments to

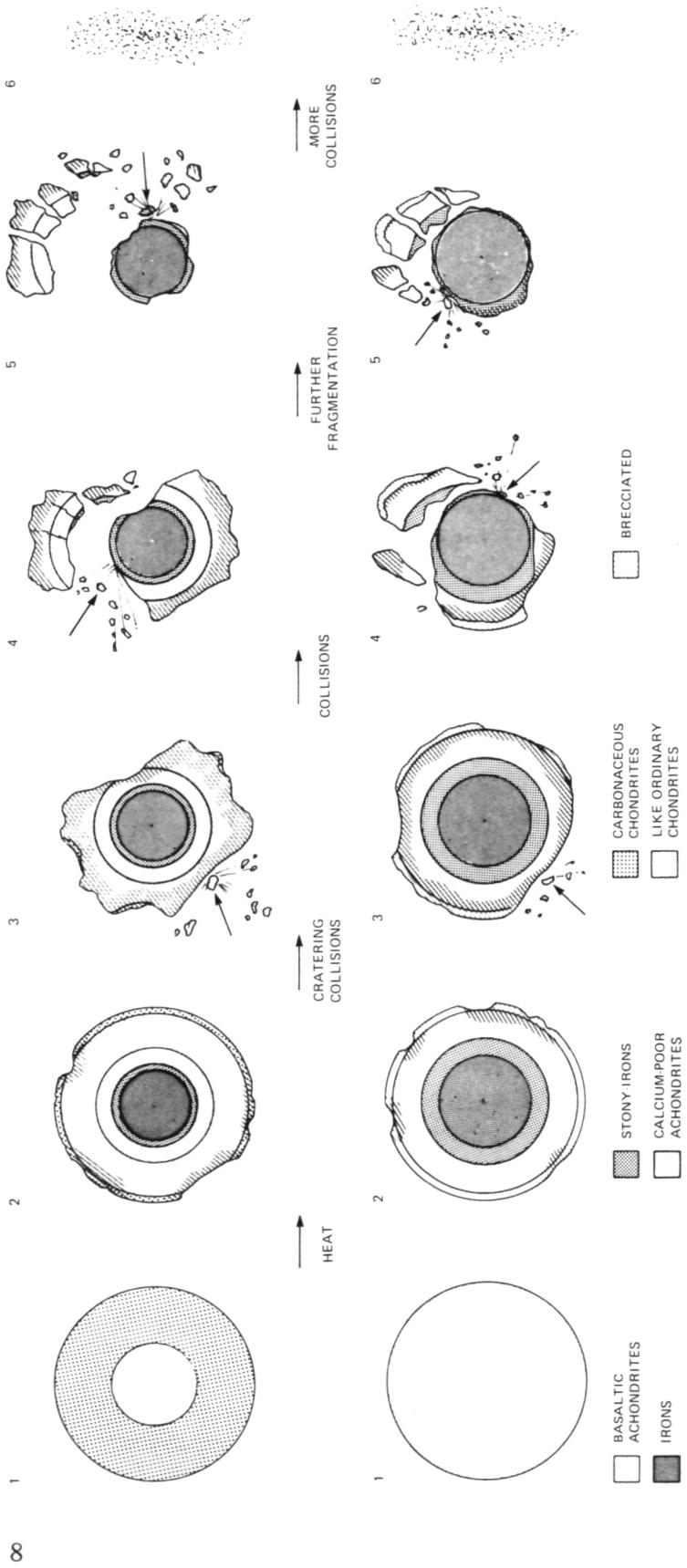


Figure 1.2. Asteroids and meteorites may be related in one of two ways. Both sequences show the evolution of a large asteroid perhaps 650 km in diameter. Either sequence could explain the mixture of meteorites that fall on the Earth and the observed distribution of the asteroidal spectra. The basic model for both parent bodies is one proposed by John A. Wood of the Center for Astrophysics of the Harvard College Observatory and the Smithsonian Astrophysical Observatory. Depending on the process by which the center of each parent body heated up, the bodies would have differentiated to different degrees. In the top sequence the original object (1), which may have resembled the present-day asteroid Ceres, heats up and differentiates into an object (2) that may have been much like the asteroid Pallas. Collisions break the body into large chunks (3); the brecciated outer layers could have resembled the asteroid Nausikau; after further fragmentation, it could have been the source of chondrite meteorites. The core (5) may have resembled the asteroid Dembowska. Other pieces of the original mantle could have yielded calcium-poor achondrites and unbrecciated chondrites; one of the larger pieces could have resembled the asteroid Toro. In the bottom sequence, similar events involving a parent body of different composition that was formed earlier or closer to the Sun yields similar types of asteroids and meteorites. After the parent body had thoroughly heated up and then differentiated, it may have resembled the asteroid Vesta; its crust could have been a source of basaltic achondrites. Further fragmentation of the smallest objects in both sequences would have created interplanetary dust (6) that would have eventually spiraled inward toward the Sun and been consumed. From "The Nature of Asteroids," by C. R. Chapman. Copyright © 1975 by Scientific American, Inc. All rights reserved.

Earth are not equally efficient for all asteroids, however, so many other asteroidal rocks must await examination from spacecraft (Wetherill 1974). Meanwhile much progress is being made from ground-based astronomical observations in determining which asteroids may be the parent bodies of the known meteorite classes (Fig. 1.2). Once this is accomplished, the "main limitation of the record from meteorites" described by Bunch in Chapter 6 will be overcome and meteoritical analyses will have much more well-defined implications.

The asteroids apparently were interrupted in the process of accreting into a planet by the collisional effects and orbital perturbations of nearby Jupiter (Safronov 1972; Kaula and Biegelisen 1975; Weidenschilling 1975). The subsequent evolution of these innumerable asteroidal planetesimals has been controlled by collisional fragmentation and by the orbital dynamical removal of asteroids from resonances with Jupiter and from the vicinity of Mars. The asteroidal dust produced by comminution in the belt has been redistributed by radiation processes such as the Poynting-Robertson effect and may have accumulated on Mars or other planets. The asteroids physically ejected from the belt have cratered the surfaces of Mars, the Moon, Earth, Mercury, and presumably Venus and some outer planetary satellites. A detailed understanding of the source bodies for planetary cratering, both during the heavy bombardment period about 4 aeons ago and subsequently, may enable us to establish an alternative standard for absolute geologic time; so far, absolute chronologies have been established through radiogenic decay only for bodies from which we have rock samples to date (the Earth, the Moon, and some unidentified meteorite parent-bodies). Although the asteroids have certainly been responsible for some planetary cratering, an as yet undetermined fraction of craters have been created by impact by comets and other hypothesized populations of planetesimals, especially from the vicinity of Uranus and Neptune (Wetherill 1975).

Beyond the implications inherent in small bodies for nebular condensation, planetary

accretion, and early planetary cratering, the asteroids are interesting to study in their own right. Standard geologic techniques, ranging from stratigraphic and structural studies of a whole body down to detailed analysis of samples, may reveal an interesting interplay of processes peculiar to small bodies in this transition zone in the solar system between rocky and ice-rich composition. The meteorites suggest that many individual asteroids may exhibit compositional layering, ranging from relatively modest thermal metamorphism with depth to complete melting and differentiation. Detailed geologic studies should be able to demonstrate conclusively what combination of the proposed sources of heat (short-lived radionuclides, T-Tauri solar wind, solar Hayashi phase, collisional heating, etc.) has predominated in the asteroid belt.

Thus the geologic rationale for studying small bodies in the solar system lies in their great promise (1) for elucidating conditions in the early solar system, beginning with the condensation of the solar nebula, and (2) for understanding the nature of the processes that may have predominantly shaped the original surfaces of the terrestrial planets, both compositionally (from infall of asteroidal and cometary dust) and topographically (from impact cratering and basin formation).

PLANETARY GEOLOGY APPROACH

In the preceding sections, we have attempted to show the relation of planetary geology to the three main goals set before the space science community by the National Academy of Sciences in 1966. The question becomes one of how to meet those objectives and what the approach of planetary geology is to be (Fig. 1.3). In the geological exploration of the solar system, certain basic questions are asked, regardless of the planet. These questions relate to the following: (1) the present geologic state of the planet, (2) how the present state of the planet differs from its past state, and (3) how the present and past geologic conditions of the planet differ from those of other planets.

| | EARTH-BASED | FLYBY | ORBITER | LANDER | SAMPLE RETURN | MANNED |
|---------------------------------|-------------|-------|---------|--------|---------------|--------|
| MOON | ● | X | X | X | X | X |
| MARS | ● | X | X | ● | | |
| VENUS | ● | X | X | X | | |
| MERCURY | ● | X | | | | |
| OUTER PLANETS AND SATELLITES | ● | X | | | | |
| ASTEROIDS | ● | | | | | |

X = COMPLETED ● = IN PROGRESS

Figure 1.3. General strategy for exploring the solar-system progress from reconnaissance missions to progressively complex missions, and from the inner planets to the outer planets. It is unlikely that this exploration matrix will be completed in the twentieth century, particularly in regard to manned missions.

Present Geologic State

The present state of a planet concerns knowledge of its surface morphology, the composition of the surface and interior, the physical nature of the interior, and whether or not there are active processes, such as volcanism. These topics obviously are interrelated and knowledge of one topic often provides clues to the others. Surface morphology involves the identification of the type and distribution of individual landforms, such as basins, mountain chains, and volcanoes, as well as structural elements such as fault scarps. The presence of these features often signals specific geologic processes. For example, the identification of obvious volcanoes, as was done on Mars, would indicate not only the effect of volcanic processes on surface morphology, but would also indicate melting in the interior, thus placing constraints on various

models of the planetary interior and giving some indication of geochemical differentiation. Observations of “variable features” that can be related to specific landforms, such as the wind-blown streaks on Mars, provide clues to dynamic surface processes. An ability to determine surface compositions and to map their distribution as has been partly accomplished for the Moon can lead to decisions concerning planetary chemical differentiation. Sometimes morphologic features may be difficult to reconcile with current atmospheric conditions. This may reveal past variations in those conditions.

Geologic Evolution of Planets

One of the primary goals in geology is to derive the geological history involving the sequence of events and processes from the time of formation of a planet to its present state. Specific questions in this derivation include:

1. How have the previous geologic processes such as volcanism, faulting, and erosion differed from recent processes in terms of type or magnitude, or both? For example, were there periods when impact cratering dominated?
2. Is there an imprint of earlier processes on the present surface morphology?
3. If the planet appears to be geochemically differentiated on a planetary scale, when and why did differentiation occur?
4. What has been the evolution of the planetary interior?
5. How has the atmosphere evolved?

These questions are extremely difficult and, in some cases, probably impossible to answer. However, geochemical data, geological mapping, and analyses of specific surface processes such as impact cratering can lead toward such answers.

Comparative Planetology

Comparative planetology is the study of the differences and similarities exhibited among the planets. Questions involving comparative planetology tend to be answered and re-asked with each newly observed planet. The last 12 years

have seen a tremendous expansion of knowledge of the inner solar system through successful lunar and planetary missions in the form of flybys, orbiters, and landers. We are beginning to see Earth at one end of the scale of the inner planets as an extremely dynamic planet with a thin, rigid lithosphere that slides over a plastic asthenosphere. At the other end of the scale is the Moon with its very thick static lithosphere. Mars seems more Earth-like while Mercury shares many attributes with the Moon. Venus, because of its dense cloud cover and hostile surface environment, remains virtually unexplored; less is known about its present state and evolution as

a solid body than any other inner planet, although a preserved cratering record is indicated by Earth-based radar images.

The chapters that follow address the major subdivisions of planetary geology and discuss means for obtaining data relevant to those subdivisions. It is not the intention here, however, to discuss the advantages and disadvantages of the various instruments that could be employed on planetary missions, but rather to present the types of measurements that should be made and their relevance in understanding planetary objects.

REFERENCES CITED

Adams, J. B., Conel, J. E., Dunne, J. A., Fanale, F., Holstrom, G. B., and Loomis, A. A., 1967a. A strategy for scientific exploration of the terrestrial planets. *Rev. Geophys.* 7:623-661.

Adams, J. B., Conel, J. E., Dunne, J. A., Fanale, F., Holstrom, G. B., and Loomis, A. A., 1967b. Scientific strategy, objectives and experiments for Mariner missions from 1970 through 1973. Calif. Inst. Tech. Jet Propulsion Lab. Doc. 900-94 (internal document).

Allen, C. W., 1963. *Astrophysical Quantities*. 2nd ed., London: The Athone Press, Univ. of London.

Anon., 1966. Space research — Directions for the future. Natl. Acad. Sci.-Natl. Research Council Pub. 1403.

Chapman, C. R., 1974. Asteroid size distribution: Implications for the origin of stony-iron and iron meteorites. *Geophys. Res. Ltrs.* 1:341-344.

Chapman, C. R., 1975. The nature of asteroids. *Sci. Am.* 232(1):24-33.

Chapman, C. R., and Davis, D. R., 1975. Asteroid collisional evolution: Evidence for a much larger early population. *Science* 190:553-556.

Kaula, W. M., and Bigeleisen, P. E., 1975. Early scattering by Jupiter and its collision effects in the terrestrial zone. *Icarus* 25:18-33.

McCord, T. B., Adams, J. B., and Johnson, T. V., 1970. Asteroid Vesta: Spectral reflectivity and compositional implications. *Science* 168:1445-1447.

McCord, T. B., and Gaffey, M. J., 1974. Asteroids: Surface composition from reflection spectroscopy. *Science* 186:352-355.

Safronov, V. S., 1972. *Evolution of the protoplanetary cloud and formation of the Earth and the planets*. Jerusalem: Keter Press (Israel Program for Scientific Translations).

Sagan, C., 1975. The solar system. *Sci. Am.* 233(3):22-31.

Weidenschilling, S. J., 1975. Mass loss from the region of Mars and the asteroid belt. *Icarus* 26:361-366.

Wetherill, G. W., 1974. Solar system sources of meteorites and large meteoroids. *Ann. Rev. Earth Planet. Sci.* 2:303-331.

Wetherill, G. W., 1976. Late heavy bombardment of the moon and terrestrial planets. Proc. Sixth Lunar Science Conf. *Geochim. Cosmochim. Acta*, Suppl. 6, 2:1539-1559.

Page Intentionally Left Blank

CHAPTER II.

STRATIGRAPHY AND STRUCTURAL GEOLOGY

M. H. Carr, D. E. Wilhelms, R. Greeley, and J. E. Guest

PRINCIPLES

Introduction

Stratigraphy is the study of the characteristics, distribution, and chronologic succession of rocks. Structural geology is the study of rock deformation and the features that result from it. Both stratigraphy and structural geology contribute to the determination of the mode of a planet's formation and the history of internal and external processes that have operated upon it. Planets are not static and unchanging; rather their constituents constantly are rearranged to form new rocks with new distributions. Changes may result from endogenic processes such as volcanism and tectonism or from exogenic mechanisms such as meteoroid impact and erosion. Whatever the processes, a partial record of successive events is preserved in the surface rocks and structures and can be reconstructed to a considerable extent by appropriate stratigraphic and structural analysis.

The significance of any event on a planetary surface is clearest if viewed in its historical context. If, for example, volcanism ceased soon after the planet formed, then the implications for thermal history and the composition of the near-surface layers are far different from those applying if volcanic processes are still active. Stratigraphy is largely concerned with placing rock units in historical sequence and relating

them to specific events in the past. This historical perspective distinguishes stratigraphy and the discipline of geology in general from most other sciences. Geometric relations between rock units are the principal source of the data used to derive relative sequences of events.

Derivation of the sequence of events involved in the evolution of a planet's surface leads to the establishment of a generalized geological time scale. Such a time scale was established for Earth only after years of detailed field work. A simplified time scale has been derived for the Moon and a preliminary time scale is currently being derived for Mars and Mercury. These time scales provide a framework for placement of rock units in terms of age. Rocks are "age-dated" in two ways — relative ages and "absolute" ages. Relative ages indicate the sequence of rock units, that is, one unit is older or younger than another unit, without any indication of how much older or younger. Absolute ages are based on the actual time (expressed in years) since the rock unit was formed and are derived from various radiogenic dating methods. On Earth, sequences of events are correlated from one area to another by means of fossils, lithologies, isotopic dates, etc., thus providing a global time-stratigraphy. In the absence of these tools, other methods of time correlation must be found. On the Moon, geometrical methods have been highly successful in establishing a global stratigraphy, largely because the ejecta blankets and

secondary impact deposits of basins are extensive marker horizons in relation to which other units can be dated. These extensive deposits appear to have formed in a geological instant. In some cases, lunar stratigraphic units have been dated so that the geological time scale is partly keyed to radiogenic time. Extensive marker horizons may be absent or are more difficult to detect on planets where degradational processes are constantly modifying the surface, as on Mars.

Heterogeneities of Planetary Surfaces — A Stratigraphic Problem

The immediate goal of stratigraphy and structural geology is to reduce the enormous complexity of a planetary surface to comprehensible proportions by dividing the near-surface rocks into units and mapping their distribution and attitude. The complexity of Earth's crust is well known, but the complex nature of other bodies in the solar system has been less widely appreciated. The Moon was the first extraterrestrial body whose surface was studied in any detail and the heterogeneity of even this relatively small, inactive body is now well established. The basic distinction between mare and terra has long been recognized. Gilbert (1893) pointed out the existence of a blanket of material surrounding Mare Imbrium. Several additional widespread terra units, such as the Orientale blanket (McCauley 1967) and the Cayley Formation (Morris and Wilhelms 1967) were later recognized and mapped from telescopic observations. Subtle differences in the color and reflectivity showed that the maria also are heterogeneous and susceptible to subdivision (Whitaker 1966; McCord and Johnson 1969). The return of lunar samples not only confirmed these earlier inferences but also revealed widespread differences that were previously unsuspected and demonstrated that the Moon has had a complex geologic history recorded in the surface rocks. Stratigraphic studies are therefore essential for orderly scientific lunar exploration. Stratigraphic techniques used on the Moon, as originally derived from Earth-based telescopic data

(Shoemaker and Hackman 1962; Wilhelms 1970) are now well-established.

Although considerable information about the Moon can be derived from Earth-based telescopic observations, this is not true of the other terrestrial planets. Telescopic observations of Mars and Mercury show faint surface markings resulting from albedo contrasts, but this information is of limited geologic value. Mariner 10 data have shown Mercury to be similar in many ways to the Moon. Geologic mapping (Trask and Guest 1975) shows that the oldest surfaces are densely cratered and contain large impact basins. Basin formation was followed by emplacement of large tracts of plains materials on at least one face of Mercury. These plains materials are considered by many workers to be in part volcanic and analogous to maria (Strom *et al.* 1975). The youngest units are fresh impact craters, the freshest of which have ray patterns. Mercury therefore shows heterogeneities similar to those of the Moon and must also be studied by stratigraphic methods. The similarity between the Moon and Mercury has allowed direct use of the lunar techniques so that an appreciation of mercurian stratigraphy was achieved very soon after the planet's exploration (Trask and Guest 1975).

Mars (Carr *et al.* 1973) has had a much more complicated history than the Moon. Internal activity appears to have continued for a longer time, and the surface has been modified by erosional processes that have operated in different ways in different parts of the planet. It is clear that, to be significant, chemical measurements on the surface of Mars depend strongly on our understanding of the stratigraphy and geomorphology. This is important because, not only has igneous differentiation occurred on Mars, but also the processes of weathering, erosion, and transport have probably modified the chemistry of deposited sedimentary rocks. The stratigraphic techniques evolved for use on the Moon are currently being used to unravel the history of the martian surface, although allowances have had to be made for the different processes that have operated there.

The surface of Venus is little understood at the present time; however, radar data show sur-

face relief including large craters and canyons. Results from Venera 8 (Vinogradov *et al.* 1973) indicate that, in the landing area of Venera 8, the uranium, thorium, and potassium contents of surface materials are high, implying the presence of highly differentiated rocks. Again the significance of this is somewhat limited by the lack of knowledge of the degree of heterogeneity of the venusian surface, and landers such as the Venera spacecraft could provide information that would be of much greater value if the information could be set in a regional context derived from stratigraphic studies.

The Jovian satellites, Europa, Io, Ganymede, and Callisto, show surface markings that would, based on our present experience with planetary bodies, cause us to expect that they, too, will exhibit chemical heterogeneities; the same may be true for some of the other outer planet satellites.

Thus, for all solid surface planets for which data are available, surfaces are composed of heterogeneous rock units and are amenable to geologic mapping. Planetary surfaces must be interpreted in light of the internal and external geologic processes that resulted in their present configuration. Stratigraphic studies will indicate which of these two broad classes of geologic processes have operated and the relations between them. If the planet has experienced internal activity, then chemical fractionation is likely and the significance of the surface chemical data will depend directly on the local nature of the internal zonation, and on the regional geology as learned from stratigraphic studies. For a planet that has had no internal activity, stratigraphic studies will be limited only to determining the nature of the planet's external environment throughout its history.

Geologic Framework

Because of planetary heterogeneity, the stratigraphy and structure of a planet are fundamental in that they provide a framework within which other measurements may be evaluated. Chemical analyses, heat-flow measurements, seismic profiles, etc., can be planned and inter-

preted far better if the geologic setting of the place of measurement is known. Without such knowledge, chemical and physical measurements can be misleading. For example, a chemical analysis of a salt dome in Texas does not give the composition of the surface of the Earth in the south-central part of the United States, nor can a value for heat flow in Yellowstone National Park be extrapolated to the whole of the western United States. Although these are extreme cases, they serve to show that the geological context is critically important to the interpretation of geochemical and geophysical data. This is so fundamental to geologic thinking that to state it appears trite and obvious, yet it is a concept that has found only slow acceptance with regard to the exploration of other bodies in the solar system. Without geologic analysis, the surface of the planet is in danger of being treated as a homogeneous unit, a chemical analysis is likely to be considered as indicative of the chemistry of the whole surface, and the mineralogy of one part might be considered as the mineralogy of the whole. Such interpretations are demonstrably nonsensical for the Earth, Moon, and Mars, and there is little reason to think that they will prove any more valid on other planets. Only the smallest asteroids and satellites possibly have been inert and homogeneous throughout their history. Measurements made on the surface of a planet must be interpreted in light of the local stratigraphy; to do otherwise could lead to gross errors of interpretation.

Methodology

Because of dependence on remote-sensing data, the techniques of stratigraphic and structural analysis of the planets differ from those normally used on the Earth. Terrestrial rock units are defined on the basis of lithology, chemistry, and mineralogy as seen in outcrop or derived in the laboratory. Their relative ages are determined largely by relations observed in vertical sections or in outcrop patterns; correlations are made by comparison of successions of lithologies, by tracing specific horizons on the ground, by diagnostic fossils, and by isotopic

dating. In general, the stratigrapher proceeds from particular observations seen as specific outcrops to a synthesis of general patterns. The reverse procedure must be used in studying the geology of a remote surface. General patterns are observed from remote-sensing data, including Earth-based measurements, and the significance of the patterns is later checked and supplemented by orbital measurements and later by observations at a restricted number of ground sites.

For stratigraphic purposes, photography has been the most valuable source of remote-sensing information. Nearly all basic lunar stratigraphic units and relations have been determined by superposition and intersection relations, and a Moon-wide chronology has been established by dating units with respect to the extensive marker horizons. Variations in gravity, emissivity in the infrared spectrum, radar reflectivity, and color have had value largely as interpretative aids; they have not had similar value in defining and mapping units because they have not been determined with the same linear resolution as visual black and white images.

Stratigraphic techniques are successful on the lunar surface because the external environment does not vary regionally. In contrast, the surface of the Earth is greatly affected by external conditions produced by its diverse climate and vegetation. As a result, a specific rock unit may exhibit a variety of surface properties. On the Moon, the principal external process modifying the surface is bombardment by meteoroids and, because this is essentially isotropic, differences in properties of the lunar surface must reflect, along with age of exposure, differences in the properties intrinsic to the surface materials. Similar conditions appear to hold for Mercury. On Mars and Venus, regional atmospheric erosion may affect surface forms, and geologic interpretation must be made in light of these atmospheric processes and their regional variations.

Once the geologic units have been defined and their relations to one another determined, the results can be portrayed on geologic maps. Although the geology of the Moon, Mars, and Mercury has been interpreted largely from

studies of surface landforms and albedo, the resulting maps are truly geologic, for they incorporate the three basic ingredients of a geologic map: rock units, age relations, and geometric position. They are quite distinct from topographic maps, terrain maps, and physiographic maps, which are concerned only with surface form. Even where the surface topography is strongly controlled by surficial effects, as it is by wind action on Mars, we can still make geologic maps *sensu stricto*, providing we first understand the nature of the surficial processes that have operated. Unless the surface topography is controlled entirely by surficial effects, we can expect to make geologic maps of several additional planets and satellites.

Complete characterization of the rock units recognized from the images will require measurements on the surface. Thus a major goal for landed experiments is to determine the mineralogy and petrology of rocks at the landing site. The choice of a site for establishing ground control clearly will be important. In the initial stages of exploration, sites should be on geologic units that have wide distribution and in areas where there is a minimum stratigraphic ambiguity. Apollos 11, 12, and 14 were so sited on the Moon. Subsequently, as on Apollos 15, 16, and 17, sites were chosen to clarify specific scientific problems, such as presence of volcanic activity, or they were in areas where vertical sections are accessible, as near a fresh crater or steep vertical wall. In late stages of exploration, surface units and their mutual relations will be most efficiently studied using some kind of traversing vehicle, such as the Apollo 15, 16, and 17 Lunar Roving Vehicles.

Rigorous geological mapping of extraterrestrial surfaces was first demonstrated by Shoemaker and Hackmann (1962) for the Moon. Their work continued with systematic mapping of the Moon through the U.S. Geological Survey and led to the completion of 1:1,000,000 scale maps for the entire nearside (Fig. 2.1), numerous large-scale lunar site maps, and a composite 1:5,000,000-scale geologic map of the nearside (Wilhelms and McCauley 1971). Lunar geological mapping continues on the 1:250,000 series

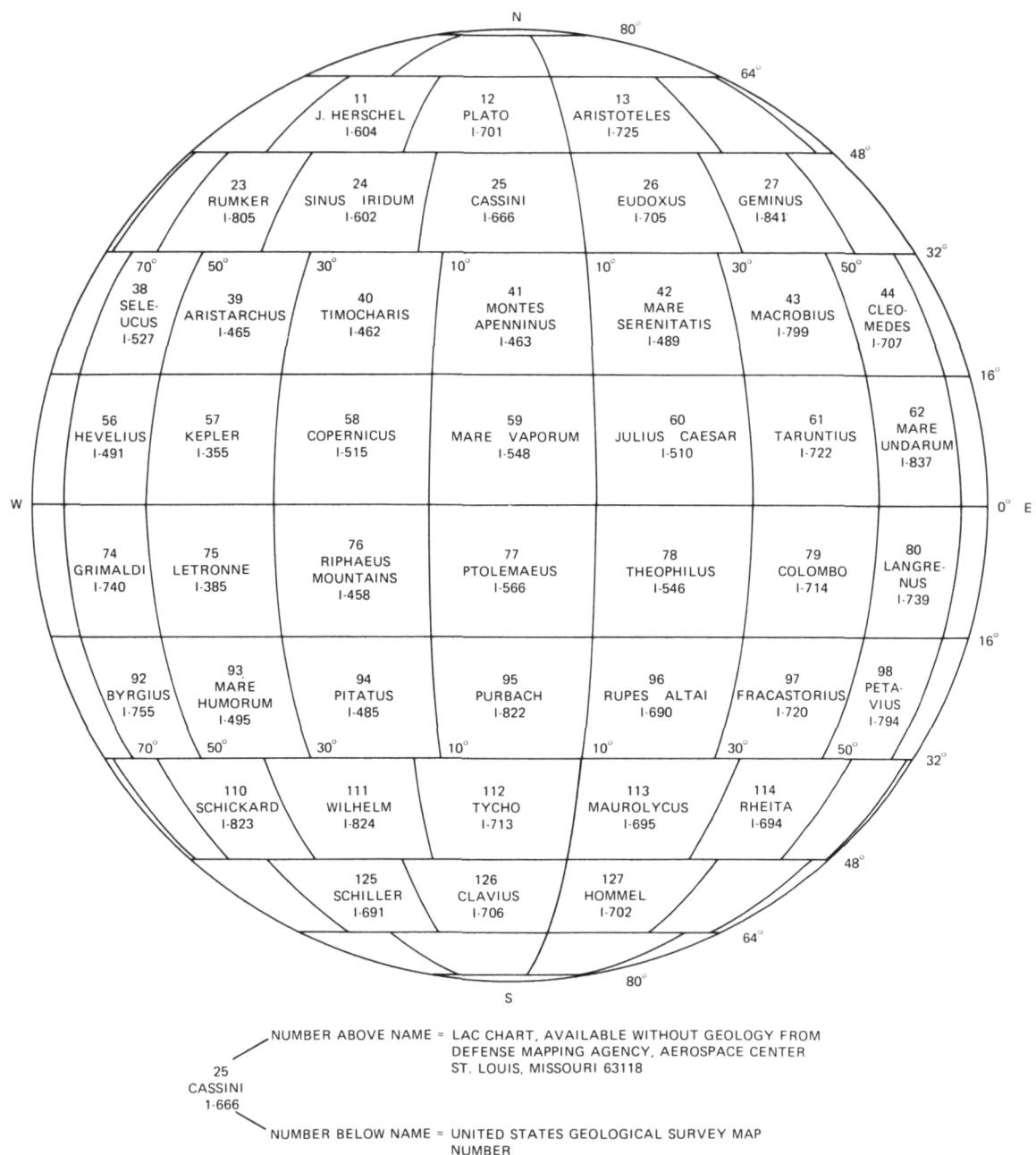


Figure 2.1. Index to 1:1,000,000-scale geological mapping of the lunar nearside, published by U.S. Geological Survey.

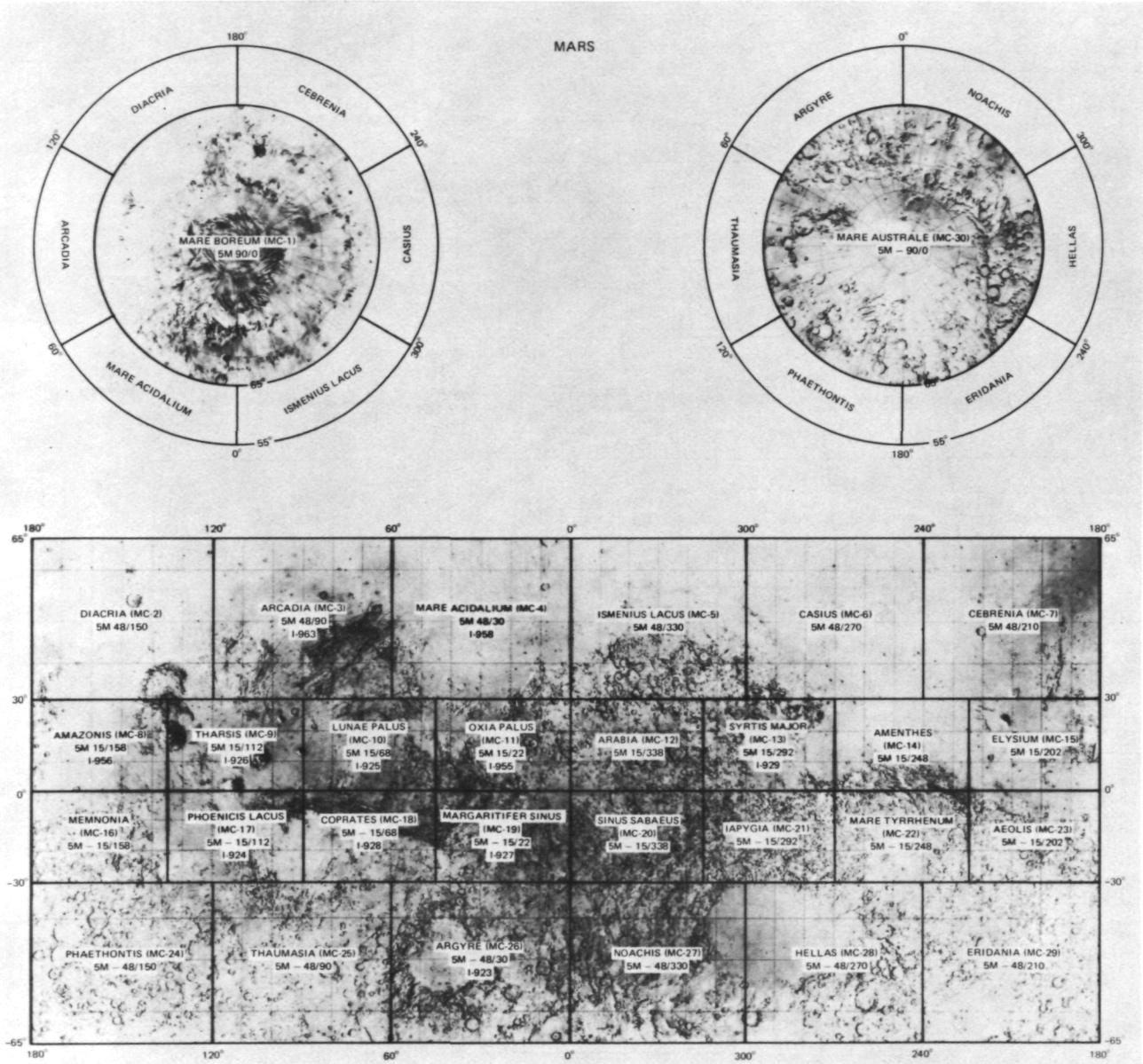


Figure 2.2. Index to 1:5,000,000-scale geological mapping of Mars, published by U.S. Geological Survey.

maps that were prepared from Apollo orbital photography and on a 1:5,000,000 scale for the entire Moon.

The Mars geologic mapping program was initiated in the early 1970's by the NASA Office of Planetary Programs to generate a complete set of 1:5,000,000 geological maps for Mars.

Using Mariner 9 images, geologists from the U.S. Geological Survey, universities, and NASA field centers have nearly completed the mapping program (Fig. 2.2). In addition, several large-scale geological maps have been generated for various Viking and Soviet landing sites on Mars.

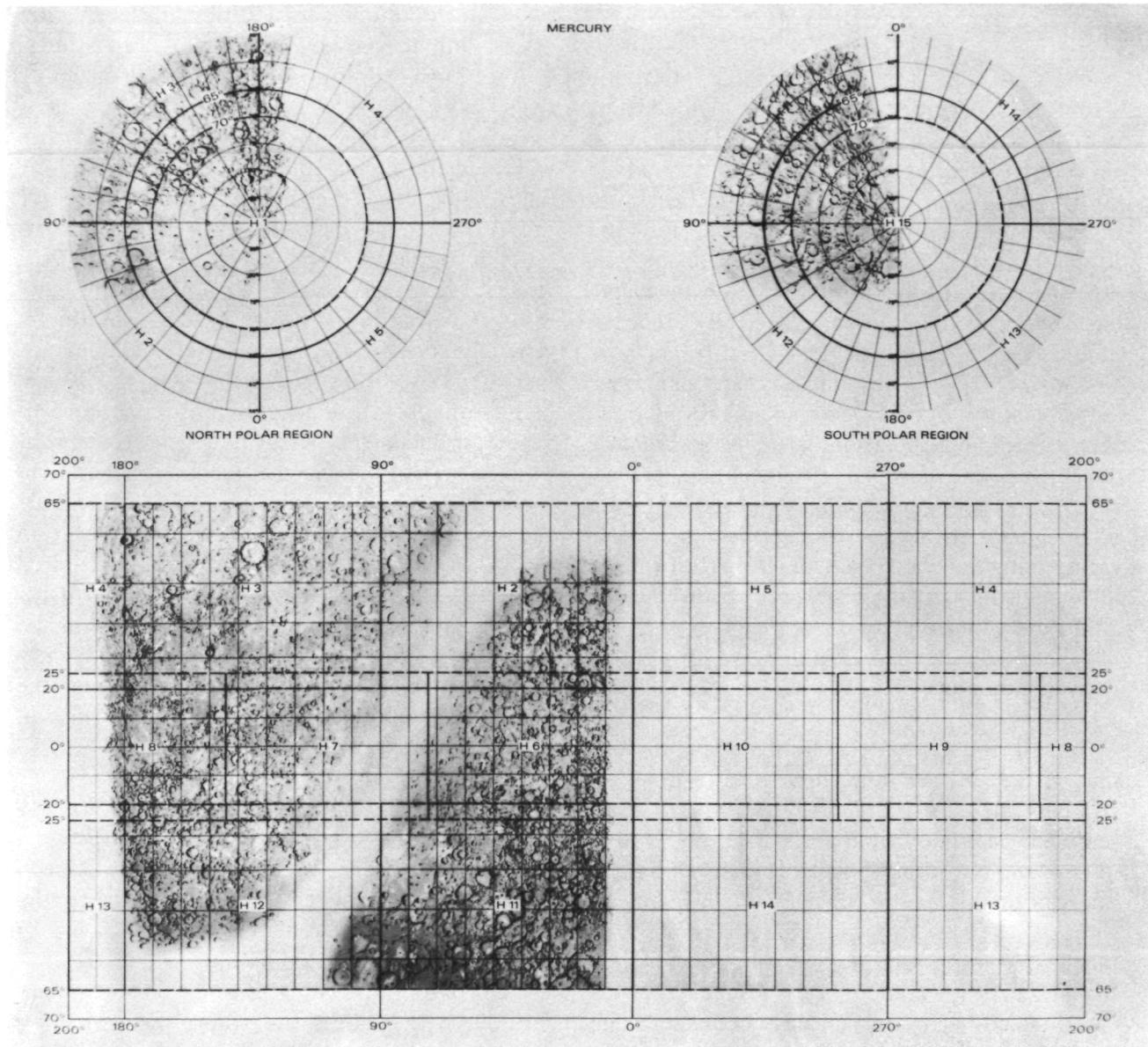


Figure 2.3. Index to 1:5,000,000-scale geological mapping of Mercury published by U.S. Geological Survey.

On the same pattern as the systematic Mars mapping, the Mercury geologic mapping program began in 1975 using Mariner 10 images. Geologic mapping at a scale of 1:500,000 will be completed for about 55 percent of the planet (Fig. 2.3).

Map scales and required accuracies vary with their intended use (landing site maps vs. regional maps, etc.), the planet under consideration, and the available images. The generation of cartographic products (see Ch. V) requires close working relations among cartographers, geologists, and other planetologists.

DATA REQUIREMENTS

The first stage in the geologic exploration of a planet is to identify, characterize, and outline the distribution of the major geologic units exposed at its surface. In the second stage, field information must be obtained at critical locations in the form of chemical, mineralogical, and petrological data from landers. Sufficient data must be obtained in the first stage so that landing sites can be chosen effectively and the obtained data related to a regional framework. No matter how many landers are deployed, determination of the geologic environment at each site and interpolation between the sites will depend on knowledge of the local and regional geology. This is most readily achieved by means of images, with other remote sensing data providing supplementary information. Normally, imaging will be in the visible spectrum, but circumstances may dictate other wavelengths.

Image Parameters

Photographs, or the more general term images, are at the same time both simple and complex pieces of data. They are simple in the sense that anyone can appreciate and have some understanding of the scene recorded by the image. For this reason, imaging experiments are valuable in conveying the excitement of planetary exploration to the general public. At the same time, however, images (especially when electronically processed) may be exceedingly complex and require great caution and understanding for proper interpretation. The various aspects of images as "links in the communication system" are discussed in detail by Davies and Murray (1971). Among the more important factors in image acquisition are resolution, illumination, color, and stereoscopic coverage; polarimetry is less important.

Spatial resolution— The value of spatial resolution is easily understood (Fig. 2.4). One of the goals of planetary image analysis is the determination of surface processes — higher resolution provides more information upon which to

base interpretations. In addition, different processes tend to produce different scale features, so that high resolution permits observation of the effects of a wider range of processes. For example, on Mars the dark markings within craters were identifiable as wind-blown dune fields only from the high-resolution frames of the "B" camera of Mariner 9. Similarly, the flow lobes, lava channels, and levees, indicative of low viscosity lava, are also visible only in the high-resolution frames. Although better resolution is generally beneficial, the cost effectiveness of increased resolution is variable. On the Moon, increasing resolution beyond approximately 5 to 10 m produces very little new geologic insight because variations on the regolith rather than the underlying rock units control the fine-scale topography. The value-function for resolution must be determined from previously acquired reconnaissance images and will be different for each planet. In practice, resolution will be restricted by the areal coverage required, space-craft performance, terrain characteristics of the planet, and so forth. Choice for each mission must be made individually.

Illumination— Images are sources of information for two basic properties of the surface of a planet — relief and reflectivity (albedo). Albedo differences are best seen at or near vertical illumination, whereas relief is best observed at lower angles of illumination where relative change of brightness for a given change in surface slope is greatest and where shadows make relief more visible (Fig. 2.4). The desirability of low illumination is true only in general and must be balanced against camera performance. Lower illumination angles mean lower absolute brightness, longer exposures, and increased shadows. On planets with atmospheres, morning and evening haze may actually lower contrast and discriminability of landforms. The choice of illumination will depend on the terrain characteristics and the camera sensitivity. With digital imaging systems, contrast enhancement techniques can do much to bring out surface detail so that choice of illumination is less critical. Generally, if enhancement procedures are to be

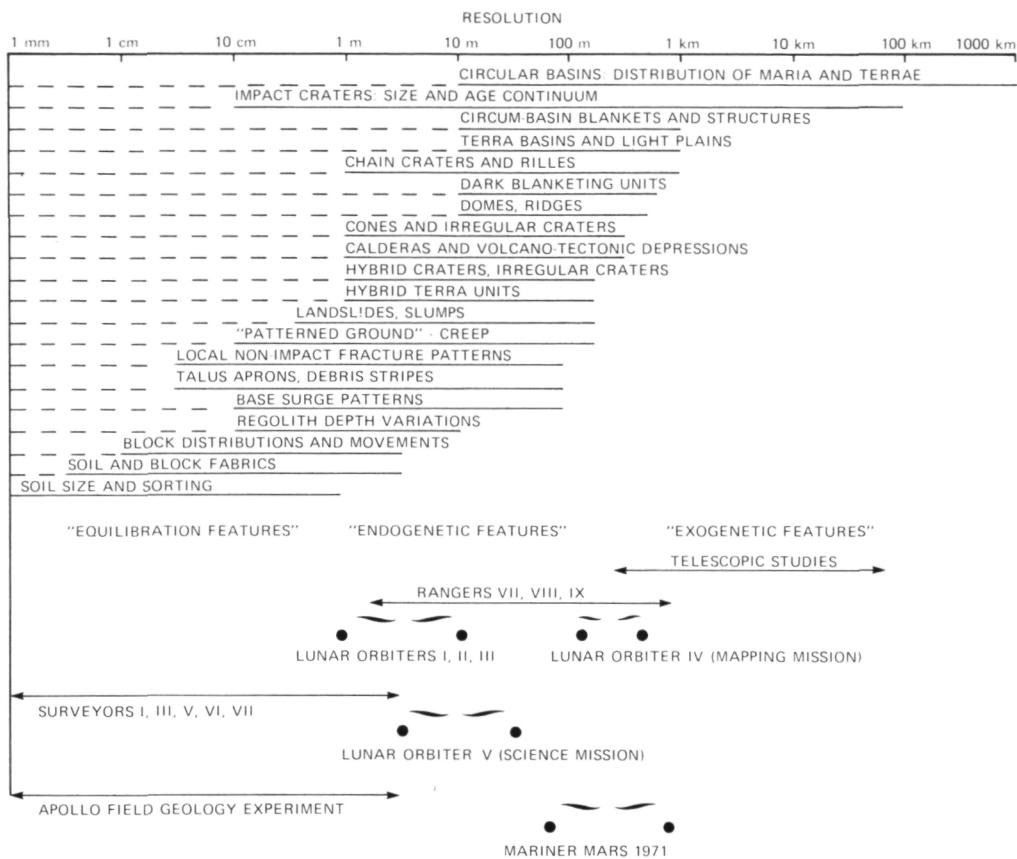


Figure 2.4. Resolution required to detect various lunar features and structures. Results from missions to Mars and Mercury show that each planet appears to have unique requirements for image resolution, dependent upon the processes that have been dominant through its geological history (from Masursky *et al.*, 1970).

applied, illumination angles up to 45° can be tolerated for landform discrimination.

Color— Color has enormous potential as an aid in geologic interpretation. Telescopic multiband photometry of the Moon has permitted the discrimination of a large number of units on the lunar surface, thereby allowing confident extrapolation of the sample site results to other areas. Much important information about rock composition can be obtained in this way. The multiband capability of ERTS transformed the experiment from one of passing geologic interest to one of substantial geologic potential. Image processing techniques to capitalize on multiband capability have advanced greatly over the last few years, mainly because of the ERTS applications. These factors have brought about an increased awareness of the value of color as part of any imaging experiment.

Color can be used in a variety of ways. First, it can be used as an enhancement tool; for example, atmospheric phenomena can be enhanced or suppressed, as can albedo variations. Second, color variations, being indicative of compositional and mineralogical variation, can be emphasized. Its main value over other geochemical techniques is that chemistry and mineralogical differences at the surface are portrayed at the same scale as the visual image and therefore can be correlated directly with topographic information.

Color has been downplayed in planetary imaging experiments up to now, and for good reasons: the greater the number of color bands, the greater are the telemetry requirements. Most imaging experiments are telemetry-limited, and it must be decided whether one wants, say, three-color imaging or three times the area

covered. In the early stage of exploration, areal coverage will and should take priority over colorimetry. As coverage increases, however, the balance should move toward color.

Another factor has worked against color: all cameras flown to planetary bodies other than the Moon have been vidicon systems, and vidicon systems flown thus far are notoriously poor photometers. Not only does the sensitivity vary over the field of view on the vidicon screen, but the light-transfer characteristics change with time, invalidating much of the ground-based calibration data. If color is to become a significant factor in an imaging experiment, then either vidicons cannot be used or a major breakthrough is required in vidicon technology, as discussed by McCord and Frankston (1975).

Stereoscopic coverage— As for color, the advantages of stereo coverage are initially less important than areal coverage; but as more monoscopic coverage is acquired, the relative value of stereo increases. Limited stereo samples of selected features should, however, have a high priority early in the exploration program. Once a representative coverage of a variety of features has been acquired with stereo coverage, color should take precedence, providing that an imaging system is being used that is photometrically credible.

Polarimetry— The usefulness of photography through polarizing filters has been marginal for the Moon and the Earth, and even for experimental surfaces under laboratory conditions. The present state of knowledge and techniques dictates that polarimetry should not be attempted at the expense of other potentially more useful techniques.

Types of Missions

Detailed comparisons of flight hardware lie outside the scope of this report. Davies and Murray (1971) and Murray *et al.* (1971) provide comparisons of various imaging systems. Thus far, all United States planetary missions have used vidicon cameras. Film was used in the Lunar

Orbiter missions and in subsequent manned flights, but it has yet to be used on a nonlunar mission. Complexity (size, weight, power), lifetime of film and chemicals, vulnerability of film to radiation, and poor enhancement capability have militated against the use of film, despite its capability of providing wide-area coverage at high resolution. However, the main disadvantages of vidicons that have been flown thus far are that they have small format, low sensitivity, poor geometry (image distortion), and poor photometry.

Recent vidicons are substantially improved in geometry, sensitivity, and photometric calibration. Recent developments in the field of solid-state cameras hold great promise for future imaging systems. Monolithic phototransistor arrays at least as large as 500×500 appear feasible. Although existing charge-coupled devices (CCD) have low resolution, high-density arrays are being developed. The inherent advantages are low power, low maintenance costs, and long lifetimes. From a scientific standpoint, they have the additional advantages of stable geometry and excellent photometry. Such devices could thus provide true multiband capability at imaging resolutions, thus adding a new dimension to imaging experiments. Clearly, development of CCD's should be vigorously pursued.

Flyby missions— The initial images of planets have been obtained from flybys, missions that have been of varying utility. Mariners 6 and 7 (Mars 1969) were reconnaissance in nature and no systematic coverage was expected; rather the purpose largely was to determine crustal style and to identify general problems. These missions gave an imbalanced view of the planet geologically because of the limited coverage resulting from a low data return rate. For optimum utility, an early reconnaissance mission should strive for images at a wide range of scales and broad coverage at moderate resolution. The ideal plan, achieved by Mariner 10 by use of a high data rate and long-focus telescopes (Murray *et al.* 1971), results in photographs of the whole disk with scattered photographs of higher resolution

located within the regional frames. This has two advantages over a plan in which all photographs are taken at the highest resolution: first, the broad geography of the body is established and, second, the nature of the terrain at different scales is determined. The second advantage is important as there is no way of knowing in advance the optimum scales for obtaining information on landforms. Systematic stratigraphic studies are possible from the Mariner 10 fly-by-type of photography; however, orbital coverage is required for a complete global stratigraphy.

Orbital missions— Orbital imaging is the single most important source of geologic information for planetary bodies. Acquisition of images from orbit of Mars by Mariner 9 transformed our concepts of the planet. One can only speculate on the bizarre interpretations that could have resulted from orbital geochemical measurements and geophysical data on Mars without the supporting images. Orbital images provide the information base against which all other data are compared and interpreted. Its usefulness depends mainly on resolution, areal coverage, and continuity. Other factors such as stereo overlap, photometric fidelity, color, and polarization are of secondary importance in the early stages of exploration, but gain in importance as exploration progresses.

Orbital images also provide the opportunity to acquire repetitive data over selected regions of the planet. For example, during the Mariner 9 mission, many regions on Mars were imaged repetitively, and variable features, such as dark streaks associated with craters, were observed to change in size and shape with time, indicating the presence of active surface processes.

Landing missions (E. Morris)— Although not strictly a technique for stratigraphic and structural analysis, lander imaging (Figs. 2.5 and 2.6) is discussed here because of its close ties with orbiter imaging. A landed imaging system should have the capability of several resolutions: a low-resolution, wide-angle mode for an overall view of the surface and a high-resolution, narrow-

angle mode for detailed examination. In addition, color and infrared capability would provide information on the composition, oxidation, and weathering states of surface materials. A landed imaging system serves three main geologic functions: (1) to examine in detail the lithology, texture, and attitude of the surface rocks at the microscale and mesoscale, (2) to determine topography in the vicinity of the spacecraft, and (3) to provide a setting for other experiments on the spacecraft.

Lithology, texture, and attitude of surface rocks: One of the main tasks of a landed imaging system is the detailed examination of the lithology, texture, and structure of surface rocks to determine the processes most active in the formation of the surface. A texture of interlocking crystal forms may indicate a crystallization stage during rock formation, either as a result of derivation from a melt or as a result of post-depositional recrystallization. Euhedral phenocrysts may point to a magmatic origin, rounded grains to a sedimentary origin. However, even when samples are available in hand, grain textures can be difficult to interpret. Textures of this type are generally visible at 1-mm resolution, and high-resolution imaging systems can achieve this resolution without additional optics. The ability to discern these fine-scale textures will depend largely on lighting, and repeated imaging of the spacecraft environment at different lighting conditions will bring out many observable features. A stereoscopic imaging system, however, is far superior to a monoscopic system for detecting and measuring fine-scale textures and features.

The gross features of rocks in place are also related to origin. Observations of bedding, cross-cutting relations, faults, and folds are all necessary to understand sequential relations, yet they cannot be made adequately from a stationary vehicle or from an orbiter. Some mobility is required because of the scale of the features involved and because a sufficiently large number of relations must be observed to confidently interpret them. Therefore, an automated mobile vehicle to test the heterogeneity inferred from the orbital images and to sample a variety of

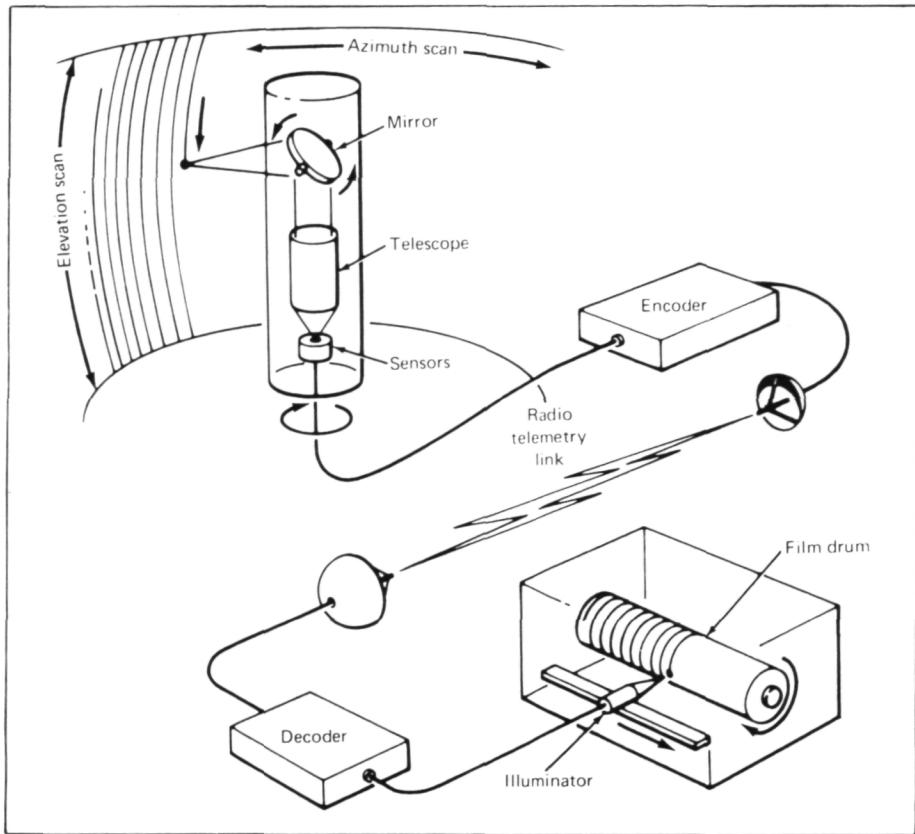


Figure 2.5. Diagram of the Viking lander imaging system. Onboard camera scans the scene and transmits the image in much the same way as the facsimile process transmits newspaper pictures.

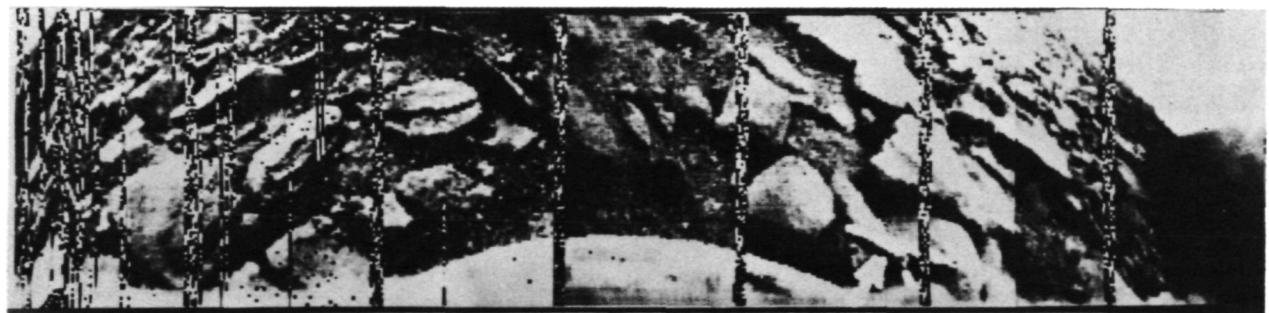


Figure 2.6. Image of the surface of Venus taken by the Soviet Venera 9 lander in late 1975. Image system is similar to the U.S. Viking lander facsimile camera. Before this image was acquired, virtually nothing was known of the Venusian surface. Several geological facts are evident: (1) the surface is rough and bouldery, (2) rock size is bimodal (i.e., large blocks and rather fine grains), (3) rocks are subrounded, although some are also angular, (4) many blocks appear to be platy, and (5) there appear to be different rock textures. These and other observations combined with data from other experiments yield a great deal of information on surface processes and local geological history (Tass photograph and caption information, courtesy M. Malin, Jet Propulsion Laboratory).

geologic units for chemical and physical study must be considered for advanced missions.

Surface topography: The conditions necessary for deriving topographic data for planetary surfaces are given in Chapter V. The topography of any planetary surface is likely to be changing as a result of destructive and constructive processes; evidence of the character of the processes will be preserved in the landforms. All features, regardless of origin, will be modified by impact and surface processes, and their state of preservation will indicate something about the rate of landform development. Detailed photography of the surface therefore provides basic information on the rate and nature of the fine-scale surface modifying processes. Topographic maps of the surface, constructed from data returned by a lander photographic system, provide means to determine quantitatively the changes of surface features and also provide accurate bases for plotting and assembling the scientific data.

Analytical support: An important function of a lander imaging system is to provide support for other analytical instruments and experiments. The geologic environment around the spacecraft is potentially very diverse. Rocks of various types may be present both in the regolith and in the bedrock. An imaging system provides the essential means of monitoring the surface being analyzed. When used in conjunction with a surface sampler, the imaging system will greatly improve the versatility of many analytical instruments onboard inasmuch as it changes the sampling from a random process to one of intelligent control and selection. This applies to geological, biological, and engineering instruments. Imaging system support for meteorological experiments on planets with atmospheres is provided by observations of the formation of clouds, movements of dust storms, and particle transport by wind. An imaging experiment also can provide an estimation of aerosol particle distribution from observations of the solar aureole, brightness of celestial objects, and distribution of sky brightness. Astronomical measurements can be made using an imaging system from the

surface of a planetary body by observing satellites and other celestial objects, and the axis of rotation of the planet can be determined from these observations. Support for biological experiments could be provided by searching for visible evidence of life and for fossils.

Radar images (G. Schaber)— Venus is completely enveloped by clouds so its surface is obscured from conventional imaging systems. The only way to obtain images of the surface of Venus, or any other planet with a thick cloud cover, is to use a part of the electromagnetic spectrum with wavelengths capable of penetrating the clouds. In practice, the long wavelengths characteristic of radar systems offer this ability, and several types of radar imaging systems have been developed toward that end. Limited Earth-based radar images of the surface of Venus have been obtained, as illustrated and discussed in Chapter VI.

Side-looking airborne radar (SLAR) measures the reflectivity of the scattering medium as a function of position and, from this information, constructs an image of the surface. SLAR was first developed for military applications and is currently being used in various Earth resources programs (Fig. 2.7). The application of radar imaging systems in the exploration of Venus has been discussed in several reports (MacKay *et al.* 1973; Colin *et al.* 1975, and others). In spacecraft use, the transmitter and receiver are both onboard and a signal is transmitted to the target surface at a fixed angle. The reflected signal varies in intensity, largely according to the relation between surface slopes and the spacecraft. The signal is measured as a function of track position, azimuth, and range. The positional and intensity information is combined to construct a radar image that strongly resembles an optical image of a surface under oblique illumination. Use of synthetic aperture techniques permits high-resolution images to be formed from orbital altitudes without the need of impossibly large antennas. The images can be interpreted by standard optical techniques provided the appropriate images are obtained. Several techniques unique

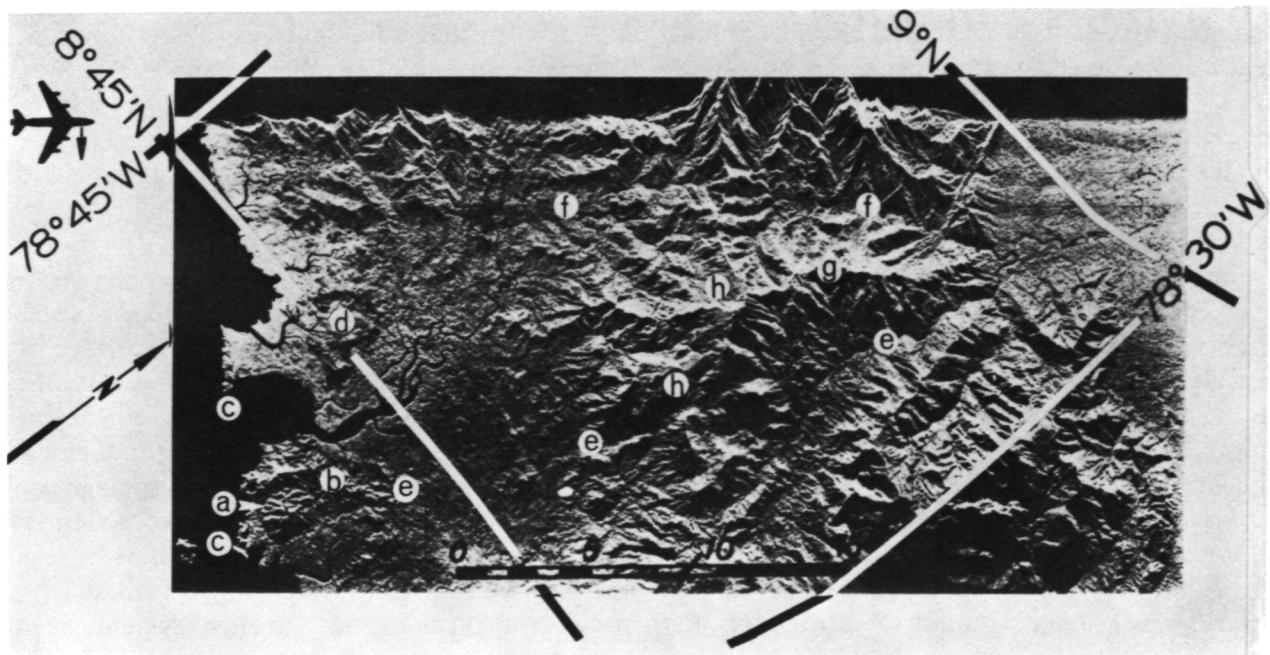


Figure 2.7. Radar image of the Chiman coastal area, adjoining the Gulf of Panama, showing primarily the Chiman igneous plug (a) and remnants of its caldera ring (b). Also displayed are nearby coastal igneous dikes (c); another caldera (d) and an adjacent small plug. This image also shows the Chiman fault (e), the West Ipeti fault (f), and the West Ipeti block (g), together with its component Huevito syncline (h). Radar layover and foreshortening in the near range (near the top) are illustrated. (from Wing 1971)

to radar image processing can also be applied, such as those developed for the Lunar Sounder radar experiment (Phillips *et al.* 1973).

In addition to images, nadir surface (topographic) profiles can be obtained simultaneously with the SLAR image by use of nearly normal (0° to 45°) look-angles and the process of coherent altimetry similar to that proposed for the Pioneer-Venus radar altimeter in 1978.

Several studies (MacKay *et al.* 1973; Friedman and Rose 1973; and others) have shown that between 80 and 90 percent of the planet Venus can be covered by SLAR images at 100- to 200-m resolution and 2 to 4 percent covered at 50 m or better resolution. This can be accomplished, according to the studies, during a baseline mission of 120 days in orbit (half a venusian day).

Use of spacecraft radar-imaging systems will depend on the rate and refinement of Earth-based radar techniques. Moderate-resolution (about 2 km) radar images of Venus must be

obtained during the next 3 to 4 years and, although it appears that the terrestrial radar observatories may possibly reach this resolution level (see Ch. 6), the percentage of the planet covered at that level will be far too small to characterize the geologic processes and materials of the surface. As such, an orbital sidelooking radar mission must be seriously considered.

IMAGING STRATEGY

B. Murray

The typical sequence in planetary exploration progresses from flyby (Fig. 2.8) missions to landers, with each type of mission having characteristic capabilities. In addition to this general mission strategy, there is a more specific strategy for image acquisition that must be considered. It includes tradeoffs between areal coverage and resolution, and the advantages of full-disk images, contiguous coverage, and nested high-resolution frames.

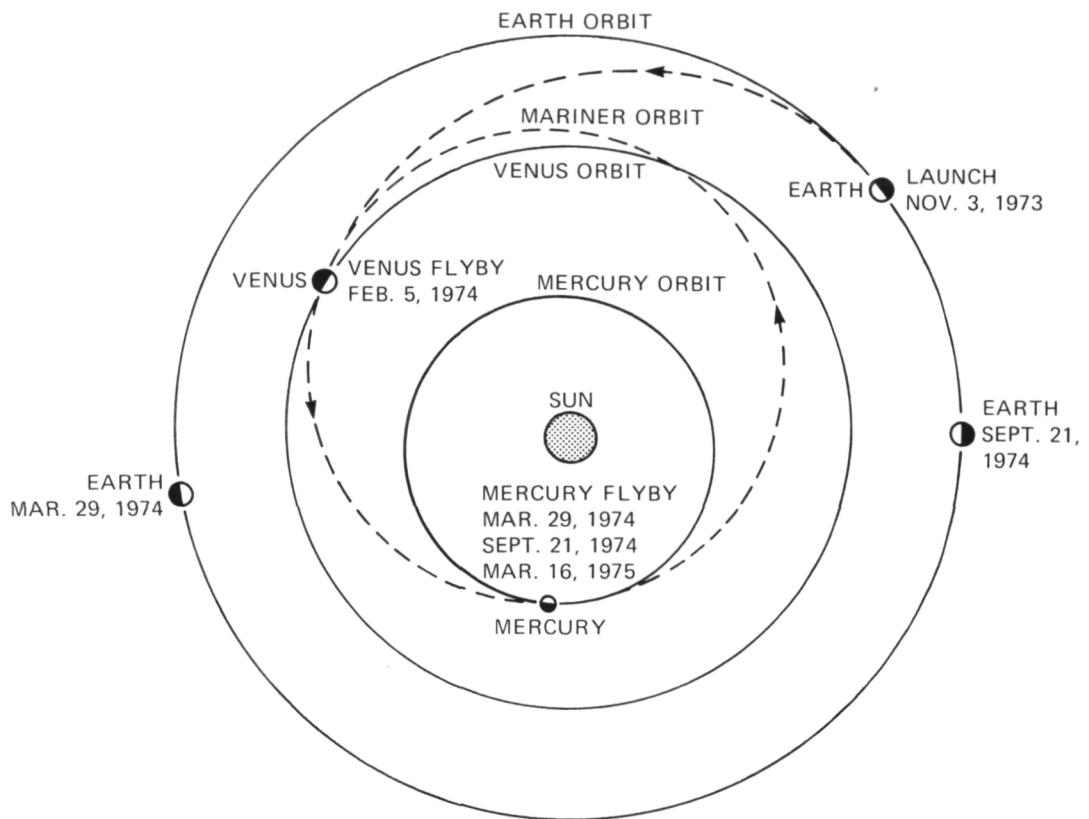


Figure 2.8. Diagram showing the trajectory of the Mariner 10 mission to Venus and Mercury. Spacecraft trajectory and orbit of Mercury was such that Mariner 10 effectively had three “flybys” of the planet, with useful data collected on each encounter.

Areal Coverage

The study of stratigraphy and structure from images hinges on recognition of regional patterns. Contiguous coverage of large areas is therefore essential. Images of restricted areas, even if of excellent quality, have limited usefulness unless the general context of the areas is known. For example, the Mariner 9 high-resolution frames (B camera), being widely scattered, cannot be used to delineate surface units or determine superposition relations. Their usefulness is restricted to characterizing units recognized at low-resolution scales (A-camera frames). The importance of areal coverage results from the fact that the distribution of a unit and its relation to adjacent units often tells more about its origin and age than does its surface morphology. For example, the fact that the Fra

Mauro Formation on the Moon occurs all the way around the Imbrium basin is far more indicative of its origin than is its surface texture. Again, very detailed photography of a particular scarp in the Southern Highlands might reveal very little about its origin, whereas wide photographic coverage could show it to be part of a system of linear features radial to Mare Imbrium and therefore related in origin to the Imbrium basin. Distribution patterns are an essential element in geologic analysis; without them, image analysis is reduced to inefficient and undisciplined exercises in comparative geomorphology.

The interplay of resolution and coverage can be demonstrated by a review of the photographic exploration of the Moon, Mars, and Mercury. In Figure 2.9a, the history of lunar photographic exploration starting from ground-based coverage preceding the space age and

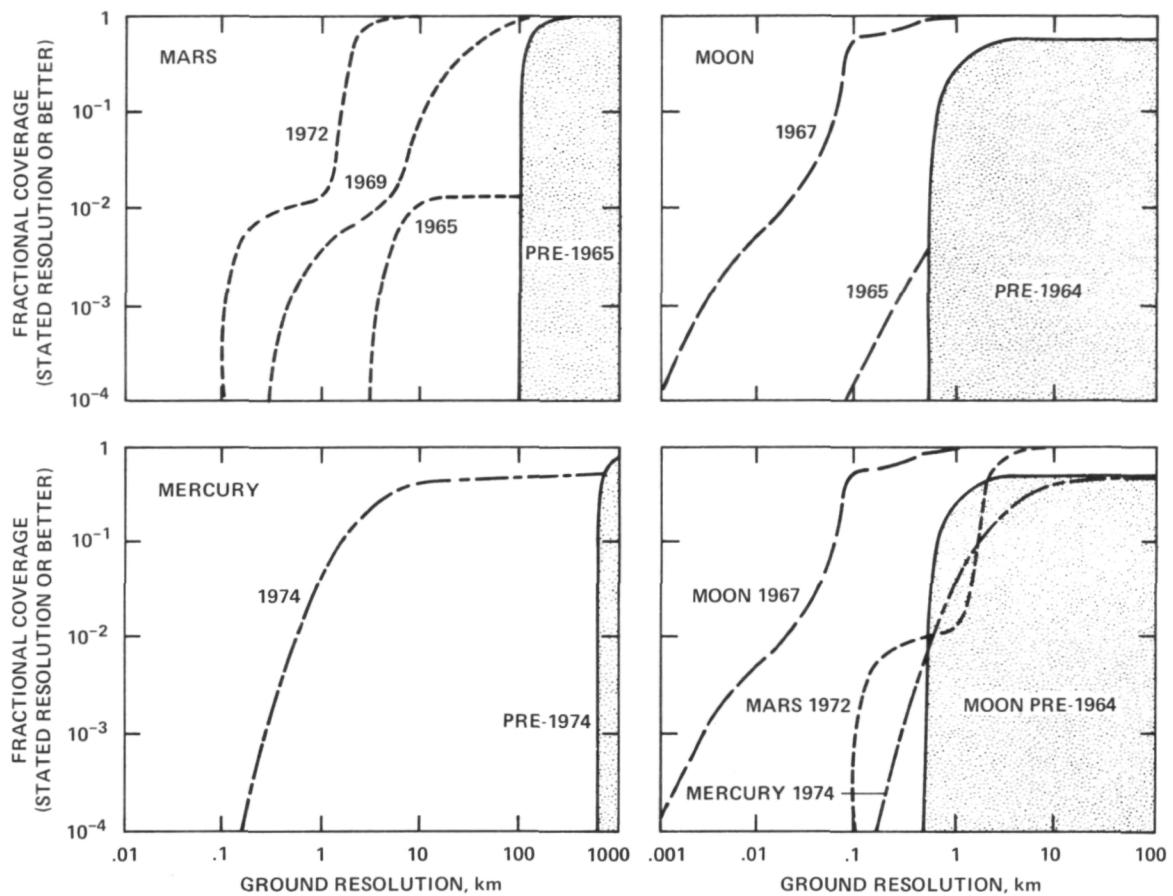


Figure 2.9. RESOLUTION VS. COVERAGE—All curves are derived by the same guidelines: (1) cumulative fractional coverage as a function of surface resolution, (2) elimination of redundant coverage of the same location, and (3) adjustment of resolution for foreshortening. (See Murray et al. 1971 and Davies and Murray 1971 for details.)

continuing through the Lunar Orbiter series is illustrated in semiquantitative form. Data returned by Lunar Orbiter IV very substantially increased our knowledge of the lunar surface, both in resolution and coverage. That increase in surface knowledge was sufficient to permit the understanding and rather complete documentation of the effects of endogenic processes as distinguished from those of external origin (such as impact). Figure 2.9b is a similar comparison done for Mars and shows the results of Mariners 4, 6, 7, and 9. In this case, the knowledge gained from Mariner 9 also permitted a significant breakthrough in the recognition of endogenic processes, as well as documentation of various erosional features. The breakthrough

came about not so much due to the increase in resolution, but because of the much greater areal coverage that included large volcanic features not shown in the earlier coverage of limited areas. Figure 2.9c illustrates the resolution and coverage of data obtained for Mercury by Mariner 10. Finally, Figure 2.9d compares the resolution and coverage data for the Moon, Mercury, and Mars. It can be seen that even after Mariner 9 and Mariner 10, we have barely attained the photographic coverage available for the Moon before the space age. For Mars, however, this has been sufficient to disclose something about endogenic processes, whereas for the Moon it was not. In addition, the Mariner 9 coverage was sufficient to unravel the puzzling

TABLE 2.1. IMAGING EXPERIMENTS FOR PLANETARY SURFACES

| Observational technique | Scientific objectives |
|-------------------------------|---|
| Full-disk images | Geometric and Bond albedo; polarization vs. phase; albedo and color markings; gross shape and radius; frost deposits; horizontal and vertical structure |
| Contiguous coverage | Evidence of impact, erosion, and internal activity; regional color and polarization; control net and maps; radii and figure of planet; slopes and heights |
| Nested high-resolution frames | Recognition of small-scale characteristics of impact, erosion, and internally caused features; atmosphere studies |

variable markings, which have to result partly from atmospheric effects and partly from local variations in the interaction of surface winds and topography. These graphs are important in gauging directly the value of various coverages and resolutions in the exploration of the large satellites in the outer portion of the solar system.

A threefold categorization is presented here which is appropriate for the observation of planetary surfaces from distant or close flyby encounters, or from an orbiter: (1) full-disk images, (2) contiguous coverage, and (3) nested high-resolution frames. An overview is presented in Table 2.1.

Full-Disk Images

The term full-disk images refers to synoptic coverage regardless of whether the illumination is nearly full-Sun, 0° phase angle, as characterizes global photography, or higher phase angle where the terminator region can be viewed.

Trajectories of flyby missions often have slant ranges (spacecraft to planet distance) on the order of $10^4 - 10^6$ km, except where an intentional close approach is possible and realized. Even a very small satellite, such as the martian satellite Phobos, can be studied meaningfully at such distances. In 1969, Phobos was photographed at a range of 1.3×10^5 km by Mariner 7. Although the image of the satellite was only about 6×8 picture elements in size, it was possible to determine that the object was elongate in the martian equatorial plane, and that it was much larger than previously supposed. Thus, it

was recognized to have a geometric albedo of 0.06, the lowest then known for any object in the solar system (Smith 1970). These inferences were confirmed in 1971 when Mariner 9 photographed Phobos from as close as 5540 km. A good deal of additional information on surface structure and history then became accessible (Pollack *et al.* 1972). However, even when satellites are barely resolvable, the Bond albedo may be determined. Due to the very limited phase excursions by the major planets observable from Earth, it is not possible to measure directly the visible phase function nor therefore the Bond albedo. Hence the radiation balance can never be ascertained directly from Earth. A most important objective of outer planet imaging (and an essential complement to infrared radiometry) is to acquire this phase function information by observation by satellites through a wide range of phase angles. Also, the variation of polarization versus phase angle may be useful in distinguishing between alternative models of surface particle composition and size, especially for the generally nonlunar satellite materials evident from ground-based results (Johnson and McCord 1970; Hansen¹).

Full-disk images also can be most useful for recognition and mapping of albedo markings, especially on the lunar-sized bodies. For satellites of the outer planets, any variation in markings with time may be indicative of a tenuous atmosphere and associated frost deposits.

¹Hansen, O. L., 1972. Ph.D. thesis (unpublished), California Institute of Technology.

Contiguous Coverage

In the photographic exploration of Mars and Mercury, it has been possible to proceed beyond the full-disk observation possible with Earth-based telescopes, and to use Mariner spacecraft to acquire groups of contiguous pictures with resolution approaching 1 km (Masursky *et al.* 1970; Murray *et al.* 1971). This capability has played a decisive role in the recognition of both exogenic and endogenic features on Mars and Mercury, and has significantly altered opinions about both planets. Accordingly, contiguous coverage approaching 1-km resolution should be a primary objective for Venus and for close encounters with lunar-sized satellites in the outer solar system. It would be expected that there will be impact craters and it will be desirable to compare their abundances and morphology and thus their histories with those on the Moon, Mercury, and Mars.

On the other hand, it would be exciting to find totally unexpected surface features as well. In this regard, the studies of Lewis (1971) suggest the possibility that within the Galilean satellites, water and ammonia could exist in a magma-like form. Thus, one cannot rule out a surface igneous activity of a type totally unfamiliar and perhaps even unimaginable to scientists whose only previous experience is with silicate magmas of the inner solar system. It is difficult to predict what kinds of features will be seen. Contiguous coverage of large surface areas is required to resolve such possibilities. Yet, if previously unrecognized surface features were found, it might be possible to infer totally different interior conditions for these objects. Also, there may be evidence of interaction with the intense magnetic fields and perhaps radiation belts that surround these planets. There is indication already from ground-based observations in the visible and infrared that these surfaces differ significantly from those of Mars and the Moon in both composition and structure. In addition to morphological studies, contiguous coverage permits study of regional color and perhaps even polarization variations.

Finally, a major application of contiguous images is the establishment of a geodetic control net, the basis for those maps whose development guides further exploration and analysis, as discussed in Chapter V.

Nested High-Resolution Frames

In addition to contiguous coverage approaching 1 km in resolution, it is generally possible during a close flyby encounter, or from an orbiter, to acquire isolated frames of up to 10 times higher resolution located within lower-resolution coverage. These have additional value, both in the detection of significant topographic features, and for certain special applications such as slope and height determination from stereo, shadow measurements, geometric parallax, photoclinometry, and photometric function studies of selected features. Finally, even tenuous atmospheres may be detected by targeted cusp and limb photography.

SUMMARY

1. Stratigraphic and structural geology studies are concerned with documenting the nature, extent, and mutual relations of surface rocks to determine the history of a planet's surface and the nature of the processes that control its evolution. Moreover, such studies provide a framework for the interpretation of surface measurements.

2. The basic data requirement is imaging, visual or radar, at a wide variety of scales and with substantial, contiguous areal coverage. The initial requirement is for global coverage at ground resolutions no less than about 1 km, then contiguous coverage of selected areas constituting approximately 10 to 20 percent of the planet and with ground resolutions no less than 50 to 100 m. These resolutions will depend to some extent on the relief and the nature of the surface. Imaging also should be obtained at high-angle illumination for albedo characterization.

3. Colorimetric, photometric, and polarization demands on the imaging system initially

should be secondary to the requirement for wide areal coverage at useful resolutions, but colorimetric characterization of the surface should follow acquisition of the initial imagery, or be done concurrently, if resolution and areal coverage are not jeopardized.

4. Once moderate resolution global coverage and higher resolution coverage of selected areas have been acquired, emphasis should shift toward remote-sensing techniques other than imaging, particularly those such as visual and

infrared spectroscopy that are sensitive to chemical and mineralogical differences.

5. Lander images and direct chemical and mineralogical determinations on the surface, or preferably on returned rocks, are essential for the interpretation of the composition and origin of units identified from orbital imagery. The stratigraphic value of the lander experiments is greatly enhanced if there is the capability of mobility or drilling or both.

REFERENCES CITED

Carr, M. H., Masursky, H., and Saunders, R. S., 1973. A generalized geologic map of Mars. *J. Geophys. Res.* 78:4031-4046.

Colin, L., Evans, L., Greeley, R., Quaide, W., Schaupp, R., Seiff, A., and Young, R., 1976. The future exploration of Venus. NASA TM X-62,450.

Davies, M. E., and Murray, B. C., 1971. *The View from Space: Photographic Exploration of the Planets*. New York and London: Columbia Univ. Press.

Gilbert, G. K., 1893. The Moon's face: a study of the origin of its features. *Philos. Soc. Washington Bull.* 12:241-292.

Johnson, T. V., and McCord, T. B., 1970. Galilean satellites: the spectral reflectivity 0.30 - 1.10 micron. *Icarus* 13:37-42.

Lewis, J. S., 1971. Satellites of the outer planets: their physical and chemical nature. *Icarus* 15:174-185.

MacKay, J. S., et al. 1973. A Preliminary Analysis of A Radar-mapping Mission to Venus. NASA TM X-2868.

Masursky, H., et al. 1970. Television experiment for Mariner Mars 1971. *Icarus* 12:10-45.

McCauley, J. F., 1967. The nature of the lunar surface as determined by systematic geologic mapping, in Runcorn, S. K., ed. *Mantles of the earth and terrestrial planets*. New York: John Wiley and Sons - Interscience Publishers, pp. 431-460.

McCord, T. B., and Frankston, M., 1975. Silicon diode array vidicons at the telescope: observational experience. *Applied Optics* 14:1437-1446.

McCord, T. B., and Johnson, T. V., 1969. Relative spectral reflectivity 0.4-1 μ of selected areas of the lunar surface. *J. Geophys. Res.* 74:4395-4401.

Morris, E. C., and Wilhelms, D. E., 1967. Geologic map of the Julius Caesar quadrangle of the Moon. U.S. Geol. Survey Map I-510.

Murray, B. C., Belton, M., Danielson, G., Davies, M. E., Kuiper, G. P., O'Leary, B. T., Soumi, V. E., and Trask, N. J., 1971. Imaging of Mercury and Venus from a Flyby. *Icarus* 15:153-173.

Phillips, R. J., et al., 1973. Apollo Lunar Sounder Experiment, Apollo 17 Preliminary Science Report. NASA SP-330, pp. 22-1 to 22-26.

Pollack, J. B., Veverka, J., Noland, M., Sagan, C., Hartmann, W. K., Duxbury, T. C., Born, G. H., Milton, D. J., and Smith, B. A., 1972. Mariner 9 television observations of Phobos and Deimos. *Icarus* 17:394-407.

Shoemaker, E. M., and Hackman, R. J., 1962. Stratigraphic basis for a lunar time scale, in Kopal, Zdenek, and Mikhailov, Z. K., eds., *The Moon - International Astronomical Union Symposium 14*. New York: Academic Press, pp. 289-300.

Smith, B. A., 1970. Phobos: Preliminary results from Mariner 7. *Science* 168:828–830.

Strom, R. G., Trask, N. J., and Guest, J. E., 1975. Tectonism and volcanism on Mercury. *J. Geophys. Res.* 80:2478–2507.

Trask, N. J., and Guest, J. E., 1975. Preliminary geologic terrain map of Mercury. *J. Geophys. Res.* 80:2461–2477.

Vinogradov, A. P., Surkov, Yu. A., and Kirnozov, F. F., 1973. The content of uranium, thorium, and potassium in the rocks of Venus as measured by Venera 8. *Icarus* 20:253–259.

Whitaker, E. A., 1966. The surface of the Moon in Hess, W. N., Menzel, D. H., and O'Keefe, J. A., eds., *The nature of the lunar surface: Proc. IAU-NASA Symposium*. Baltimore: Johns Hopkins Press, pp. 79–98.

Wilhelms, D. E., 1970. Summary of lunar stratigraphy – telescopic observations. U.S. Geol. Survey. Prof. Paper 599-F.

Wilhelms, D. E., and McCauley, J. F., 1971. Geologic map of the near side of the Moon. U.S. Geol. Survey Map I-703.

Wing, R. S., 1971. Structural analysis from radar imagery of the eastern Panamanian isthmus, Part II: *Modern Geol.* 2:75–127.

CHAPTER III. GEOCHEMISTRY

M. H. Carr and P. Toulmin III

Investigations of the elemental and mineralogical composition of a planet bear on several important questions: (1) What were the solar-system conditions under which the planet formed? (2) What processes have altered or modified the early chemical and physical state of the planet? (3) What is the present chemical and physical state of the planet? Ideally, to answer the first question, one must know the bulk starting composition of the planet, from which estimates can be made of the temperature, pressure, and composition of that part of the solar nebula from which the planetary material condensed. In practice, one cannot determine the bulk starting composition because the initial state has been modified and because we can investigate directly only the geologic unit now exposed on the surface. The problem of modification is addressed by investigating the composition and the formative processes of the various surface units, after which inferences can be made regarding the primordial material. By such methods, for example, lunar material can be shown to be depleted, on a planetary scale, in the low-temperature, volatile-rich components of the solar-system starting material. The Moon must have formed, at least in part, from material initially formed in the hotter regions of the nebula than did material making up the more volatile-rich Earth or from another body (e.g., the Earth) in which fractionation had taken place.

The investigation of the surface rock composition and processes relates more directly to the evolution of the planet (Fig. 3.1). First, analysis of surface units leads to knowledge of the actual formational process. Clearly, sampling will be critical. One cannot do this statistically on the assumption of random variability. The plan must be highly selective, guided by an assessment of which samples will yield the greatest geologic insight, and based on an understanding of the geochemical regularities and geologic relations among the various rock types of the body. We should begin by determining the mineralogic and petrologic relations in order that the processes by which the major rock units have been formed can be inferred and the relative proportions of different rock types can be deduced. Only when we have a fairly detailed understanding of the geochemical and petrogenetic processes that have operated on the planet, and the history of their action, together with a satisfactory model for the internal structure of the planet, will we be in a position to arrive at a reasonable evaluation of the chemical evolution of the planet.

The most effective strategy for the geochemical study of a planet is to proceed systematically with the mineralogical and chemical characterization of its materials in accordance with geologically determined priorities. It is insufficient merely to analyze chemically the surface rocks. To appreciate the meaning of a chemical analysis, some assessment must be made of the

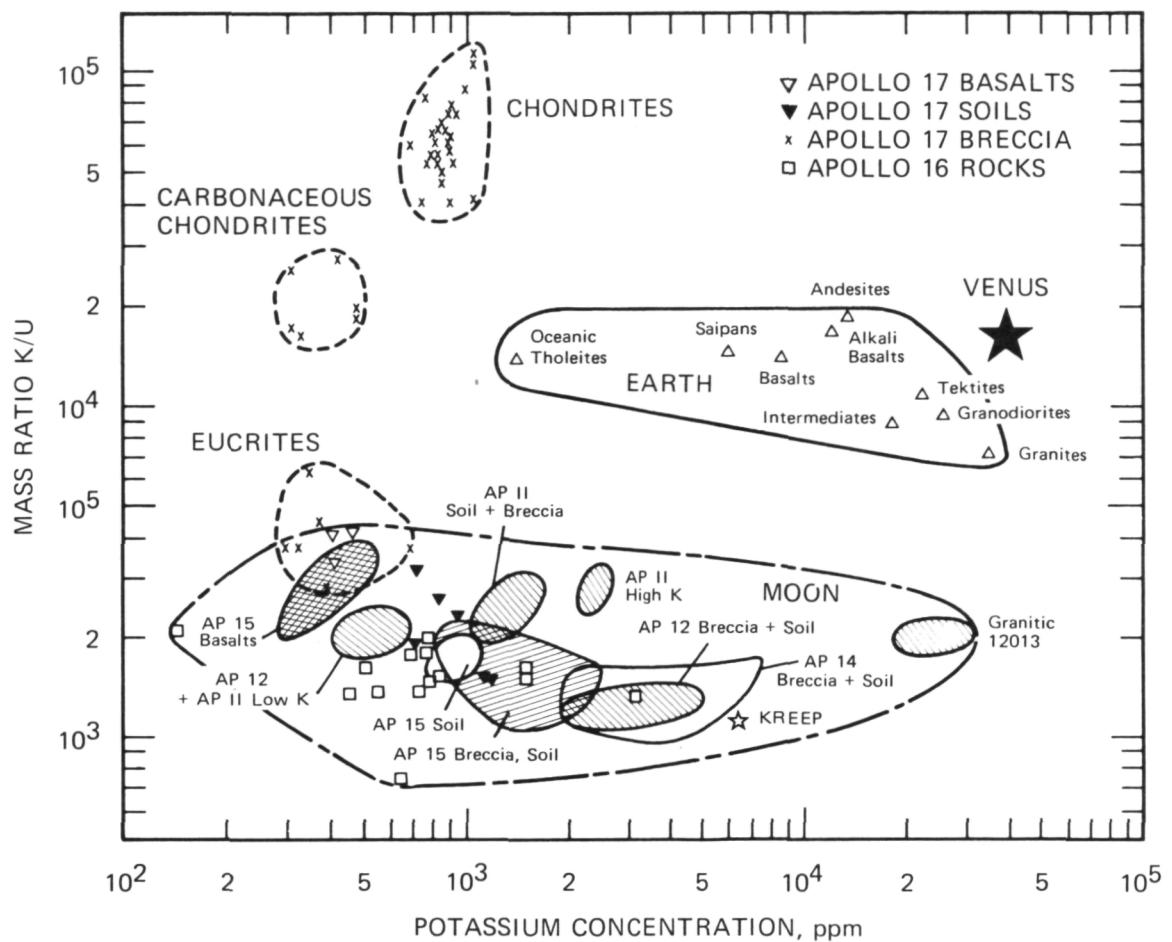


Figure 3.1. Variations in the mass ratio of K/U relative to potassium concentration (in parts per million) in lunar rocks compared with stony meteorites and terrestrial igneous rocks. Almost all rocks from these three planetary sources can be distinguished chemically on this basis. Star indicates value for Venus obtained from Soviet Venera results. (Modified from Short 1975, after O'Kelley et al. 1972.)

geologic history of the sample — what its source and mode of origin are and what processes have operated upon the sample to cause chemical fractionation.

Determination of mineralogy, texture, lithology, and other properties of the rock that might be relevant to origin is, therefore, necessary. Mineralogic composition is particularly important; the mineral assemblage reflects both major element chemistry and conditions of formation and is, therefore, far more diagnostic of rock types and formative processes than chemical composition alone. Indeed, if a choice must be made between a mineralogic and a chemical analysis, many geologists will opt for the mineral-

logic analysis because it provides much more than just chemical information. Such a choice should not be necessary, however, because mineralogical and chemical experiments can be integrated readily to provide a complete analysis. Furthermore, the choice would depend on the breadth of each analysis, the number of elements determined (and with what precision), and the sophistication of the mineralogic determination.

The exploration strategy for the planets, satellites, and asteroids must be more comprehensive than that for the Moon, whose early exploration was conducted with the expectation of early return of lunar samples to the Earth. Difficult

experiments could be deferred until a sample was available on Earth. Because the acquisition of samples from any of the planets will be far more costly and technologically more difficult than similar lunar missions, the maximum possible scientific explorations should precede sample return missions. Difficult experiments, such as age determinations, petrologic characterization, and isotopic and trace-element analyses, should not be delayed until samples are returned to Earth; the instruments necessary for performing these experiments remotely (including landers) must be developed and deployed.

The success of any geochemical exploration program will depend largely on how intelligently sites are chosen for analysis. A chemical analysis on the surface of a sand dune on Mars, for example, would not be expected to reveal much about petrologic processes. Information from orbiters and Earth-based instruments will be crucial in selecting sites. Photography is likely to be most valuable but other techniques more sensitive to chemical composition, such as multiband photometry, also should be used. The usefulness of these techniques depends not so much on the precision with which they can determine surface chemistry, but on their ability to detect differences. Because orbital measurements are important in giving areal dimension to specific surface analyses, the orbital and landing programs should be planned in concert and strategies employed such that each mission plan can be modified by information gained from the preceding program.

Acquisition sites must be chosen to provide maximum return of planetary knowledge. For this reason, careful preparatory study will be required to determine the variability of surface materials, areal geologic units, and their modes of origin. The importance of adequate documentation at sample sites cannot be overemphasized. Enormous scientific return can be expected from any returned sample, but much potential information of a truly planetary significance will be lost in the absence of a detailed geologic context.

In summary, data from a lander should include both chemical and mineralogical infor-

mation supplemented by information of petrologic importance such as texture, lithology, and structural position. The general significance of such information gathered at a restricted number of landing sites should be evaluated in light of all available geologic information, including geophysical measurements on properties of the interior, and orbital measurements of the variation in surface properties. The following sections contain more specific discussions that evaluate different methods of acquiring the necessary fundamental data.

ELEMENTAL COMPOSITION

Sampling

Some of the regional aspects of sampling programs were discussed in the previous section, but local problems are to be expected in the immediate vicinity of the spacecraft. Because of possible local variations, analytical instruments must have the capability of making measurements on any nearby object of interest and of visually monitoring such measurements. This requires that sampling tools be available and that the analyses be conducted in close conjunction with an imaging system, as was done successfully with Surveyor on the Moon.

The importance of a chemical analysis, including both minor and major elements, cannot be overemphasized. The present chemistry of a rock is the end result of a wide variety of fractionation processes that have operated on the constituent materials during the time since the solar nebula was formed. Many of these processes have left their imprint on the ratios of the different elements present. Rarely is evidence of a previous fractionation process completely destroyed by a later process. For example, all the lunar samples preserve evidence of an event early in lunar history that resulted in partial separation of volatile and nonvolatile elements. One task of the geochemist therefore is to unravel the sequence of fractionation steps using the present proportions of elements present in a rock. This task ideally requires that all elements be determined. A second task is to define the

rock type being analyzed so that petrologic arguments and inferences can be pursued. Generally, for the second task, only major elements need to be determined.

Accuracy and Precision

It serves very little purpose at this time to specify accuracies and precisions for possible chemical determinations. That determination will depend on what is technologically feasible and also on the variability in surface chemistry of the planet under consideration. Efforts to achieve extreme precision with relatively few analyses should be balanced against the desirability of many analyses to account for the natural chemical variability of the rocks present. In Table 3.1, the precisions listed are those adequate to characterize most terrestrial rock types. For even the most homogeneous terrestrial rocks, the natural chemical variability generally exceeds the listed precisions so that, to more accurately define the chemistry of the rock, multiple analyses are required as well as increased analytical precision.

Table 3.1 serves only as a guide to precisions needed to define rock types. This, however, is only one, and perhaps only a minor, function of a chemical analysis. Its more important function as a record of chemical fractionation requires that as many elements be determined as is practical, for all elements participate in a fractionation event.

The relevance of specific elemental ratios to conditions of accretion and subsequent differentiation of planetary bodies has long been recognized. In 1967, the geochemistry group at the NASA-sponsored Santa Cruz conference noted the importance of many such ratios and suggested intensive study to evaluate the usefulness and interpretive value of these and other elemental relationships. Subsequent research has but emphasized this basic message (see, e.g., the recent review by Grossman and Larimer (1974) for a discussion of chemical fractionations in the early history of the solar system). Relationships among elements that differ markedly with respect to volatility, refractoriness, oxidation-reduction behavior, crystal-chemical behavior, or other specific processes of chemical fractiona-

TABLE 3.1. PRIORITIES AND PRECISIONS FOR ELEMENT DETERMINATIONS

| First priority | Expected range, weight percent | Precision needed to define most terrestrial rocks, percent of amount present |
|------------------|--------------------------------|--|
| Na | 0.5 - 5 | 10 |
| Mg | 0.5 - 20 | 10 |
| Al | 2.5 - 20 | 10 |
| Si | 15 - 35 | 3 |
| K | 0.5 - 5 | 5 |
| Ca | 0.5 - 6.5 | 10 |
| Fe | 0.5 - 20 | 5 |
| U | --- | |
| Th | --- | |
| H ₂ O | --- | |
| CO ₂ | --- | |

Elements of secondary importance are S, Sr, Ni, Pb, Cr, and Li. Next in order of interest would be Ba, Cu, Mn, P, B, Cl, F, and Zn.

tion may reveal much about the history of a planetary body. As an example, the constancy of the ratios of volatile to nonvolatile elements in lunar materials has been used to support the argument that the Moon accreted homogeneously (Brett 1973; Philpotts *et al.* 1972; Duncan *et al.* 1973).

The cosmically abundant light elements C, O, and H are of great chemical interest as well, though the principal concern is with their molecular, rather than elemental, chemistry. In principle, independent oxygen analysis can be used to estimate the oxidation-reduction state of a sample, but in practice the ambiguities introduced by hydration and the presence of several elements of variable valence restrict it to a corroborative role. Abundances of H_2O , CO_2 , CO, and CH_4 , their modes of occurrence as simple condensed phases, clathrates, or adsorbed and occluded gases in planetary materials, as well as in planetary atmospheres, are obviously very important to both geologically and biologically oriented investigators.

Analytical Techniques

A detailed evaluation of different analytical techniques is beyond the scope of this report, and the following discussion is included more to illustrate problems than to suggest solutions. No one technique can be used to determine all the necessary major elements satisfactorily; different techniques may have to be combined if we are to have an adequate analysis. Since a mineralogical analysis is absolutely essential for understanding any chemical analysis, compatibility with phase-detecting devices is a principal criterion in evaluating the relative merits of the different techniques.

The alpha-backscatter was used successfully on the Surveyor program (Turkevich 1973). The technique gives good results for the light elements — especially O, Mg, Al, and Si, which are common in rocks most likely to be found — but it is imprecise for heavier elements. Lack of precision for the heavy elements is particularly limiting for the highly diagnostic elements — Ca, K, and Fe. However, the inclusion of an x-ray

mode would allow greater sensitivity and resolution for the heavier elements. The alpha-backscatter technique would encounter problems on Venus because of atmospheric absorption, but it might be appropriate for Mercury and Mars. Since its use in the Lunar Surveyor program, alpha-backscatter instruments have been reduced considerably in size and offer good potential for small landed spacecraft.

X-ray fluorescence is a versatile technique with a high degree of compatibility with various methods of phase analysis and one by which most elements other than those with $Z < 11$ can be examined, depending on the excitation source. The standard technique of using a high-energy primary x-ray beam to excite the sample is sensitive for all elements heavier than Al^{13} . To determine the lighter elements, advantage can be taken of the rising alpha-particle excitation function with decreasing atomic number. An alpha emitter such as Cm^{242} or Po^{210} can be used to excite the sample. The excited x-rays may be analyzed dispersively by means of goniometers or nondispersively by solid-state detectors. Continuing development of small multichannel analyzers and higher resolution detectors makes the nondispersive methods potentially very attractive because the instruments are mechanically simple and because the resolution of present detectors (<200 V) already permits unambiguous determination of adjacent elements heavier than Mg^{12} . The Viking x-ray fluorescence spectrometer makes effective use of radioactive x-ray sources and energy-dispersive, sealed gas proportional counters (Toulmin *et al.* 1973).

Neutron activation may be of value in determining elements whose determination is not possible by one of the more general techniques, or in low-level determinations of specific elements. The technique is particularly sensitive for Na, Al, Si, Ca, K, and Mn and could be combined conveniently with an x-ray fluorescence unit using the same analyzer. The only additional requirement would be some form of neutron source. The sensitivity of neutron activation depends on the particular element and on the flux available to excite the sample, but poten-

tially the technique can be used for trace-element analysis not possible by either scatter or x-ray fluorescence.

Optical emission spectrometry, the traditional method of trace-element analysis, is potentially adaptable for remote applications. The development of linear arrays of closely spaced photodetectors may have provided the key to utilization of this technique in space, but photography with subsequent photometric scanning is still a possibility. The method is sensitive to a large number of elements, it can deal with very small samples, and it has a wide range of response. Calibration to provide quantitative data may be the main problem with the technique.

Recent advances in scanning electron microscopy (SEM) and energy dispersive analysis (EDA) have considerable promise for future space missions. The SEM-EDA allows identification, observation, and analysis of individual components in the regolith, thereby permitting the analysis of many geologic units in one sample. Analytical techniques recently developed for pollution studies which involve the minimum of sample manipulation may well be applicable to planetary exploration. Small, simple, desk-top SEM's now on the market are logical prototypes for flight instruments and it appears technically feasible to develop flight instruments by straightforward modification of existing instruments.

It is evident from this brief discussion that no single technique will fulfill all geochemical needs. Emphasis should be placed on integrating various methods and exploring the commonality of different components so that a complete geochemical package can be assembled, similar in concept to the Apollo Lunar Surface Experiment Package (ALSEP) for the Moon.

ISOTOPIC COMPOSITION

R. Zartman

Any nuclear or thermodynamic process that affects the isotopic proportions of an element may be considered for isotopic composition study. Areas of interest to planetary science

include (1) radioactive decay, including age determination and heat production, (2) interaction phenomena with solar and cosmic radiation, (3) primordial isotopic abundances and heterogeneities of the original solar system, and (4) mass fractionation among stable isotopes. In this section, only mass ratio measurements that are not dependent on detecting emission from a radioactive species are considered. Gamma-ray spectroscopy (subject of a later section) offers a much more satisfactory method of determining certain radioisotopes by remote-sensing or instrument-landing spacecraft.

Perhaps the single most fundamental contribution of isotope research is in providing information about the chronology of geologic processes. Only by placing rocks into the context of both proper time sequence and absolute age can we achieve order out of the complexity of a planet's surface and arrive at some understanding of its evolution. Vastly different implications concerning the chemistry and internal conditions of a planet arise, depending on whether its surface formed 4.5 billion years ago and has remained virtually unchanged, or is still undergoing endogenetic change. The isotopes of main interest for dating long-lived events are those of $(^{40}\text{K} \rightarrow)^{40}\text{Ar}$, $(^{87}\text{Rb} \rightarrow)^{87}\text{Sr}$, and $(^{238}\text{U}, ^{235}\text{U}, \text{and } ^{232}\text{Th} \rightarrow)^{206}\text{Pb}, ^{207}\text{Pb}$, and ^{208}Pb . Several other isotopic systems have potential for dating, depending on the chemical nature of the sample, but necessary developmental work must precede their general application. The enrichment in radiogenic isotope(s) to be expected for each dating method will depend on the integrated chemical history of a particular sample. The original chemical composition of a planet and the extent of its differentiation can have a major impact on how meaningful a chronology can be established by instrument-landing spacecraft. Certainly, in such a situation, the great precision possible with elaborate chemical processing and mass spectrometry in a specialized laboratory would not be available. However, even if the required elemental and isotopic ratios were known to 10–50 percent, extremely valuable time constraints might be placed by each of the dating methods. Because

the isotopic composition also bears on the geochemical relationship of parent to daughter element, these data find additional use in understanding the petrogenetic history of a planetary body. The ability to establish broadly a time framework and to delimit gross chemical evolution would, of course, be enhanced by a careful selection of landing sites that would sample across the available stratigraphic succession of rocks.

Many nuclear transformations are induced by interaction of solar and cosmic radiation (dominantly high-energy protons with some contribution of heavier charged particles) with planetary surfaces, atmospheres, or both. In the complete absence of a protective atmosphere, as on the Moon, solar radiation will penetrate to depths of several millimeters and cosmic radiation to depths of several meters in silicate rock before its energy is spent, mainly by spallation and secondary neutron reactions. Depending on the nature of a surrounding atmosphere, this radiation may be partially, or completely, absorbed before reaching the surface. In addition to producing radioactive species with half-lives ranging from seconds to millions of years, the nuclear reactions can effect permanent changes among the stable isotopes from contributions of spallogenic and cosmogenic components. Studies of such anomalies, particularly in the noble gases, certain rare earths, and in the D/H ratio, can provide useful information about processes and mixing rates among the regolith, hydrosphere, and atmosphere.

Strong differences in bulk chemistry exist among planets and satellites of the solar system. These may reflect the conditions of condensation from the solar nebula or be caused by late events such as fission of protobodies. Isotopic distinctions also have been found recently among certain classes of meteorites, suggesting that not all such objects derive from one original homogeneous reservoir of chemical elements. If primordial isotopic variations are characteristic of the major bodies of the solar system, allowance for this nonuniformity must be included in theories of nucleosynthesis and in cosmic mixing models. At present, the isotopic effects attribut-

able to primordial heterogeneities have been small and would require very sophisticated instrumentation on a spacecraft to detect.

In contrast to the mechanisms discussed above for bringing about differences in isotopic composition by nuclear means, an important area of research is based on the thermodynamic properties of nuclides, which cause them to be mass-fractionated physiochemically. Thus, processes of evaporation, crystallization, chemical reaction, and diffusion will result in a variable distribution of isotopes of an element between phases or across a phase gradient. The fractionation effect is greatest for the lighter mass elements (D/H , $^{13}C/^{12}C$, $^{15}N/^{14}N$, $^{18}O/^{16}O$, $^{34}S/^{32}S$), which could be especially useful in the study of volatile movements, magmatic activity, geothermometry, and provenance.

The primary instrument for making isotopic composition measurements is the mass spectrometer, although certain other techniques such as neutron activation and proton backscattering may be applied to specific isotopes. Currently available flight instruments that utilize gas or sputtering ion sources allow only the crudest determination of isotopic ratios and generally have been designed instead to measure elemental abundances. Recently, efforts have concentrated on the development of a miniature ion microprobe as the single most useful chemical and isotopic measuring instrument. Again, the ion microprobe would probably find far greater application in elemental analysis, but it is hoped that broad constraints may also be made on some critical isotopic parameters.

MINERAL COMPOSITION

X-ray diffraction is a promising technique for determining mineralogy. No other technique can provide the same combination of generality, specificity, and quantitative data. X-ray diffraction patterns of polycrystalline material can be interpreted: (1) to identify, with little or no ambiguity, most phases present, (2) to determine quantitatively the composition of most solid-solution series, and (3) to determine quantitatively or semiquantitatively the modal pro-

portions of phases in aggregate. The phase compositional data can be recalculated into an elemental analysis which, though less precise than that obtained by direct methods, is nevertheless useful for most interpretive purposes. (Interpretation of an elemental analysis involves, at least implicitly, calculation of a hypothetical mineralogical composition.)

Several different diffraction techniques are possible. The most commonly used procedure in terrestrial laboratories employs a movable detector whose angular position relative to the sample and primary x-ray beam is varied to measure the angular dispersion of the diffracted x-ray energy. Although this technique has been adapted for spaceflight applications, the necessity for maintaining precise geometrical alignment is a limiting factor. Other techniques, such as position-sensitive detectors, photographic methods, and energy-dispersive diffractometry, should be thoroughly investigated.

Another technique of mineralogical determination that deserves mention is differential thermal analysis (DTA). This technique, although of limited use for systematic mineralogy, can provide very useful mineralogical data. It is particularly important in identifying clay minerals, carbonates, and other minerals that undergo low-temperature transformations. The devices are compact and lightweight and have low power and transmission requirements; therefore, they may still be feasible when a more comprehensive mineralogical experiment is not possible. Addition of effluent gas analysis, for example, by mass spectrometry, greatly enhances the scientific return of this technique.

ORGANIC CHEMISTRY

K. A. Kvenvolden

A major objective in planetary exploration should be to understand the chemistry of the organogenic elements, C, H, N, and O, the most abundant of the reactive elements in the universe. One important consequence of the interaction and evolution of compounds formed from these elements is life, and the search for extraterrestrial life is one significant part of this

objective. The ultimate goal is to know the organic chemistry of the planets (and biochemistry, if life is present). This goal includes understanding the origin, distribution, and fate of organic material in our solar system and, by extrapolation, in other regions of the cosmos. Such an understanding could also lead to an increased understanding of the origin of life on Earth and possibly elsewhere.

Organic compounds can form in several ways: (1) they can form during the early evolution of the planet and may represent material similar to that from which life arose on Earth, (2) they can be generated by interactions, promoted by natural energy sources, between constituents of planetary atmospheres and surfaces, and thus can reflect the results of past or present processes, (3) they can come from external sources such as carbonaceous meteorites, (4) they can result from biological processes and be in the form of morphological and molecular (chemical) fossils, and (5) they can be the result of living organisms such as much of the organic material observed currently on Earth. One challenge in planetary exploration will be to differentiate among these and perhaps other possibilities. Discrimination between the results of biological and abiological processes is of obvious importance.

Wherever organic material is found during planetary exploration, it should be examined in the context of the inorganic material with which it occurs. Microscopic examinations are necessary to show relationships between organic material and the associated inorganic environment; such studies will yield a visual history of the organic material and will be especially important if extant or extinct (fossil) life is present. Determinations of both the molecular and isotopic compositions of the organic material are important to fully characterize the material. For example, the various kinds of molecules, their structure, stereochemistry, and distribution provide evidence for mechanisms and processes; isotopic compositions ($^{13}\text{C}/^{12}\text{C}$, $^{15}\text{N}/^{14}\text{N}$, $^{18}\text{O}/^{16}\text{O}$, D/H) reveal where the processes have involved isotopic fractionation. Investigation of the organic chemistry of the planets should start with reconnaissance and be

followed by exploration and detailed studies.

Reconnaissance involves examination by means of orbital and probe missions utilizing spectroscopic and analytical techniques such as infrared spectrometry, mass spectrometry, and gas chromatography. Reconnaissance will be particularly effective for determinations of compositions of atmospheres. Exploration can be accomplished on planetary surfaces by remotely operated, miniaturized chemical laboratories that utilize, for example, gas chromatography-mass spectrometry, and remote biological laboratories. Detailed studies require that samples be returned to Earth for examination in laboratories where all relevant analytical tools and approaches can be used.

Because of the difficulties in designing and fabricating remotely operated laboratories that can provide unequivocal results, prime consideration should be given to returning samples to Earth. A returned sample mission is especially important for the study of Mars, although possible back-contamination problems would have to be thoroughly investigated. For the other planets and satellites, the constituents of the atmospheres of Venus, Jupiter, Saturn, and particularly Saturn's satellite, Titan, should be analyzed for organic constituents by probe or orbiter or both. The return of samples from an asteroid for study on Earth should follow.

PLANETARY VOLATILES

F. P. Fanale

Investigation of the history of solar-system volatiles should involve seeking the answer to three broad, central questions, the second of which may be subdivided into three subquestions:

1. What are the differences in *bulk* volatile content and composition among planetary objects, and how did those differences arise?

2. How has the volatile component of the material that formed each planetary object been subsequently redistributed within each object and partitioned between its interior and atmosphere?

a. What is the total inventory of degassed volatiles that has been supplied to each planetary surface environment?

b. When did the supply occur and what were the dominant energy sources responsible for the degassing?

c. What was the geochemical fate of each volatile constituent supplied?

3. What is the influence of volatile content on internal differentiation and tectonic processes?

The main points made in this section are:

1. Initial differences in bulk volatile content among planetary objects has been one of the main reasons, along with differences in mass, for divergence of the thermal and differentiation histories of planetary objects.

2. Some "small" bodies, including the outer-planet satellites and certain asteroids, have experienced intense surface differentiation and devolatization despite the fact that some of these objects cannot have differentiated exclusively as the result of decay of long-lived radio-nuclides. The devolatization history of these objects should rank high in any list of subjects for comparative planetological studies. What are the nonradiogenic or short-lived radiogenic heat sources and did they likewise affect the evolution of other planetary objects?

3. In any meaningful study of atmospheric history, it is necessary to treat the totality of all geochemical sinks for volatiles. The atmosphere and condensed surface volatiles that are in evidence on planetary objects constitute (in most cases) only a trivial and compositionally misleading fraction of their degassed volatile inventory. Comparison of degassing histories of objects can be meaningful only when sinks such as volatile-containing phases in planetary regoliths and exospheric escape have been properly assayed.

4. The rare gases constitute a partial exception to the preceding statement. They do not generally have histories severely complicated by chemical interactive effects and, with the fortunate exception of He, are essentially conserved in planetary atmospheres. Studies of radiogenic and nonradiogenic rare gas nuclides in planetary

atmospheres will allow significant statements to be made about the time history of planetary degassing as well as its overall intensity.

The study of planetary volatiles can provide profound insights into the geologic evolution of planetary objects. First, we have every reason to believe that the largely systematic variation in the uncompressed densities of planetary objects throughout the solar system reflects a systematic variation in bulk composition among planetary objects. Among nongaseous objects, this systematic variation encompasses a range from dense objects such as Mercury, rich in metal and refractory elements, to ice-rich objects such as the outer planet satellites. Thus the variation in bulk composition of planetary objects may be thought of as a variation in the ratio of volatile elements and compounds to the refractory ones, as the result of differences in the effective temperature and pressure at which nebula gas and dust equilibrated, prior to planetary accretion or loss of gas from the preplanetary nebula. In turn, these differences in temperature and pressure probably result from differences in heliocentric distance in the nebula (Cameron 1973; Lewis 1974).

Given such differences in initial bulk composition (and mass), planets will then proceed on diverse tracks of geologic evolution. For example, all other parameters being equal, and assuming steady state, the thermal gradient inside a planetary object is essentially proportional to its radius, which determines its thermal lag time. The central temperature of a planetary object may be considered to be the product of its thermal gradient and radius. Thus (all other things being equal) the ability of a planet to develop a molten interior and volcanism depends very sharply on its mass. However, volatile/refractory fractionation among planetary objects is also important. Potassium is a semivolatile element, while uranium is a refractory one. This means that refractory-rich planets such as Mercury may be depleted in K and thereby deprived of a heat source. On the other hand, this same volatile depletion process carried to the extreme may result in enrichment in U relative to major mass fraction components, thus enhancing the

ability of the object to differentiate. The latter mechanism has often been invoked to explain the Moon's apparently high level of heat flow.

The ability of an object of a given mass and given content of radionuclides to differentiate also depends on the melting point as a function of depth. For example, the Galilean satellites Ganymede and Callisto are clearly depleted both in U and K because volatiles such as H_2O ice constitute more than 2/3 of their mass. However, this also lowers their melting point so that even 1/3 or less of chondritic abundances could be sufficient to result in melting at depth in these objects (Lewis 1971). In fact, thermal models of Ganymede and Callisto suggest that these objects probably possess silicate cores, liquid water mantles hundreds of kilometers thick, and crusts consisting of either relatively pure ice or a primordial conglomerate of ice and silicates (Fanale *et al.* 1975). These objects may prove to be among the most geologically active in the solar system.

In compositional contrast, Io, the innermost Galilean satellite, appears to be an ice-free object with a bulk density of 3.5 g/cm^3 . Studies of its optical properties suggest that Io has experienced considerable degassing and defluidization, resulting in enrichment of volatile and semivolatile elements on its surface (Fanale *et al.* 1974). It seems clear, therefore, that global differentiation and devolatilization can be intense even on the small bodies of the solar system, despite the fact that most of these bodies are incapable of retaining substantial atmospheres ($>10^{-6}\text{ bar}$). Moreover, it seems safe to say that the chances of observing plate tectonic phenomena on Ganymede are as good as they are for any other object in the solar system.

Even more striking examples of differentiation and volatile loss on small bodies are provided by the asteroids. Although most objects in the asteroid belt appear to have surfaces covered by exceedingly volatile-rich carbonaceous chondritic material (Johnson and Fanale 1973) as predicted by Lewis (1972), the asteroid (4) Vesta clearly possesses a refractory-rich surface with a mineralogy similar to that of achon-

dritic meteorites (McCord *et al.* 1970; Veeder *et al.* 1975). However, there is also evidence suggesting the possible presence of hydrated compounds on its surface (Larson and Fink 1975). Since Vesta is so small that volcanism or differentiation induced by radioactive decay of U, Th, and K is virtually inconceivable, the following question arises: Are there other heat sources such as Joule heating or impact heating which could explain the fact that Vesta's surface appears to be differentiated? The larger, more important, question is, of course: Did similar processes affect the differentiation and degassing history of other objects in the solar system and, if so, how were still other objects (such as the carbonaceous chondritic asteroids) protected?

The importance of these considerations for the temporal history of differentiation and degassing of the terrestrial planets should be obvious; each energy source for differentiation has its own built-in time table. For example, if collision-induced melting occurs, it must take place at the very outset of planetary history, whereas melting induced by U, Th, and K must await accumulation of a sufficient amount of radiogenic heat per gram of material, which (even with no heat loss) requires hundreds of millions of years. Obviously, the often used anthropomorphic analogies (planets in their infancy, old age, etc.) are ill conceived when applied to systems that may have several energy sources, each with its own inherent chronology and effect on planetary evolution. Equally ill conceived is the concept that evidence for continued degassing (as a high atmospheric ^{40}Ar content or a high ^4He flux) constitutes evidence against catastrophic early degassing (Fanale 1971). On the contrary, the only connection between the two is a synergistic one: the greater the initial buried heat, the more likely it is that any given subsequent contribution of radiogenic heat will be able to produce melting at any given depth and hence degassing and volcanism.

Next we shall consider those objects that have retained substantial planetary atmospheres. In such cases, we may be tempted to hope that comparison of the mass and composition of

atmospheres and condensed phases on surfaces of planetary objects can provide insights into differences in their bulk composition and the intensity or the time history of their differentiation and degassing. While detailed exploration of planetary atmospheres and surfaces ultimately will provide information of this type, our hopes should be tempered by our experiences with the two best studied surface-volatile inventories in the solar system — those of the Earth and Mars.

It has long been recognized that almost all the Earth's degassed volatiles are contained in the oceans or chemically bound in sediments (Rubey 1951). The Earth's atmosphere represents only a small fraction of its total volatile inventory and its composition is almost completely an artifact of the combined chemical effects of biological activity, exospheric escape, etc. In the case of Mars, some investigators had suggested that the ostensible atmospheric CO_2 content (together with the CO_2 in the perennial caps) represented a substantial fraction of the total degassed martian volatile inventory and that, moreover, the entire amount could have been a late arrival, supplied by the volcanic episodes associated with the development of Olympus Mons. But this now seems unlikely in view of the fact that the thermochemical fractionation models discussed here all predict that martian material, in bulk, is more volatile-rich than terrestrial material. In addition are the following evidence and interpretations:

1. The martian atmosphere is saturated much of the time with H_2O at its base over much of the disk (Schorn *et al.* 1969).
2. Given (1), theoretical calculations show that hard-frozen H_2O permafrost is likely to occur in global lenses thickening toward the poles (Leighton and Murray 1966), and perhaps elsewhere in the regoliths as well.
3. Given (1), the regolith is theoretically expected to contain large amounts of absorbed H_2O and CO_2 (Fanale and Cannon 1974).
4. Even larger amounts of bound H_2O (up to 3 percent by weight) are suggested by infrared observations at $3\text{ }\mu\text{m}$ made from high-altitude aircraft (Houck *et al.* 1973).

5. Other visible and near-infrared reflectance spectra of Mars suggest coatings of hydrated iron oxides on regolith grains (Adams 1968).

6. The ratio of the 0.3 to 2.5 μm reflectance spectra of the light and dark regions on Mars shows residual water bands in the light regions (McCord *et al.* 1971).

7. The mid-infrared transmission spectrum of dust levitated in the great 1971 duststorm has led to tentative identification of montmorillonitic clay as a major constituent (Hunt *et al.* 1973).

8. Both hydrated iron oxides and montmorillonite have been effectively produced (from magnetite and silicates, respectively) under simulated martian conditions by UV irradiation in the absence of liquid H_2O but in the presence of adsorbed H_2O (Huguenin 1974).

9. No useful upper limit has yet been placed on the regolith's content of nitrates, carbonates, and other volatile-containing phases from the visible and infrared reflectance spectrum of Mars.

10. The unconsolidated regolith to which all the above remarks apply may extend to a depth of 1 km or even more (e.g., see Chapman *et al.* 1969; Hartman 1973).

11. Permanent surface and near-subsurface polar caps of both H_2O and CO_2 have been inferred which are much larger than the seasonal caps (Murray and Malin 1973).

12. Analysis of the exospheric energy balance deduced from Mariner 1971 orbital data suggests equal ${}^{40}\text{Ar}$ and CO_2 concentrations at 120 km altitude on Mars; this is compatible with the limits on ${}^{40}\text{Ar}$ that can be deduced from comparison of the CO_2 abundance in the atmospheric column and the total pressure at the base of that column (Levine and Riegler 1974), and is also compatible with Mars 6 planetary probe data (Sokolov 1974) suggesting a ${}^{40}\text{Ar}$ content of 35 ± 10 percent in the lower atmosphere. If true, this would require burial in the regolith of a huge inventory of chemically active volatiles. It would also provide a means to compare the total amount of (continuing) degassing, as deduced from the ${}^{40}\text{Ar}$ content with the present rate as deduced from the ${}^4\text{He}$ content, which

represents an equilibrium between current degassing and exospheric escape. Ultimately, identification of volatile-containing minerals in the regolith will provide clues as to the nature of the chemical environment on Mars' surface during its history, that is, with regard to Eh, Ph, presence or absence of an aqueous phase, etc.

Thus, the most important parameters to be measured with regard to the history of martian volatiles are (1) the abundance of rare gas nuclides in the atmosphere, and (2) the identity and abundance of volatile-containing phases in the regolith.

In a study of the temporal history of degassing of planetary objects, the Earth should not be neglected. The Earth's degassing and atmospheric history is still poorly understood. This comment applies, in particular, not only to the relative importance of catastrophic early degassing versus gradual and continuing degassing, but also to the chemical state of the early atmosphere.

Some frontiers are as follows: first, the thermal history of the Earth is still poorly understood. We need interpretation of petrologic and seismic data that will allow us to better define the K content of the Earth. We also need to establish the proportions of fossil (accretion) versus radiogenic heat in the present heat flow.

Second, we need to interpret, with detailed computational models, the elemental and isotopic composition of the rare gases in the Earth's atmosphere in terms of the transfer function of juvenile volatiles from the interior. This effort involves quantitative models to explain the degree of apparent enrichment of ${}^{129}\text{Xe}$ and fission Xe nuclides in well gas and old anorthositic rocks. These nuclides are radiogenic, they have different generation functions, and they have parent nuclides with radically different geochemical affinities. It also involves explaining the mass fractionation of nonradiogenic rare-gas nuclides between the Earth, the solar wind, and meteorites.

Third, the aeronomic problem of the ${}^4\text{He}$ balance between crustal input and exospheric escape must be solved if the differential (present) degassing rate is to be compared with the

integrated average degassing rate associated with volcanism during Earth history as deduced from the ^{40}Ar content of the atmosphere. Estimates of the rate and mechanisms of ^4He escape are currently questionable. The claimed flux of ^3He also needs to be investigated further because it is of great importance to degassing models and because fluxes from different oceanic stations appear to be very different.

Fourth, additional ways to identify the juvenile (vs. recycled) component of chemically active gases (H_2O , CO_2 , SO_2 , CH_4 , H_2S , etc.) issuing from volcanoes and thermal springs are needed. At present, only broad limits on the juvenile component can sometimes be set from isotopic data.

Thus, even for those bodies that have retained substantial atmospheres, it is unlikely that reliable information concerning the intensity or time history of planetary degassing can be obtained by directly studying chemically active gases in the atmosphere and in surface condensates. What is required is a profound understanding of the present and past role of *all* geochemical sinks for surface volatiles, from chemical recombination into volatile-containing phases in regoliths to the most ephemeral of sinks — exospheric escape. A single exception is the rare gases, the histories of which generally do not suffer from the complexities of chemical interaction or (except for ^4He) exospheric escape. The latter exception is fortunate because it allows for comparison of present and past degassing rates.

Some final recommendations concerning exploration strategy are:

1. The investigation of planetary atmospheres should be linked firmly with the investigation of chemically bound volatiles in planetary regoliths and exospheric escape phenomena so that meaningful comparisons of evolved volatile inventories among planetary objects can be made.

2. To maximize value per dollar, a balanced program is required. In such a program, theoretical studies, laboratory studies, and ground-based telescopic observations are not regarded as ancillary, but are elevated to the same status as space-craft observations, although the latter necessar-

ily must be funded at a higher level. Such a program would be problem-oriented in fact as well as by charter.

3. Foremost among planetary atmospheric measurements is a set of measurements that must be made by either probes or landers — the measurement of the abundances and elemental composition of the rare gases. No other single set of direct atmospheric measurements will provide such definitive information regarding atmospheric history.

4. There is now every indication that many outer-planet satellites and even certain asteroids have experienced differentiation and devolatilization histories as interesting as those of the terrestrial planets. Moreover, it appears unlikely that accretional heating or heating from decay of long-lived nuclides can entirely explain this phenomenon. In addition to investigation by spacecraft, support for investigating these bodies by means of ground-based, laboratory, and theoretical studies should be broadened in keeping with the spreading recognition of their fundamental importance for comparative planetological studies. The key problem to be investigated is the possibility of energy sources other than decay of long-lived radionuclides that can contribute to evolution of both large and small bodies.

WATER

D. M. Anderson

Water is now known to be widely distributed throughout the solar system. Earth-based spectroscopic observations and space probes have confirmed its existence on Mars and on several other planets and their satellites (Owen and Mason 1969; Pilcher *et al.* 1972; Houck *et al.* 1973; Fanale *et al.* 1974). Because it is essential to life, the presence or absence of water on planetary bodies is a question of great importance. Water plays many roles in various physical and chemical processes of geochemical interest as well as in biological and atmospheric phenomena. The significant questions of immediate interest deal with its modes of occurrence, distribution, and movement, its exchange with the atmosphere, and its role in the contemporary

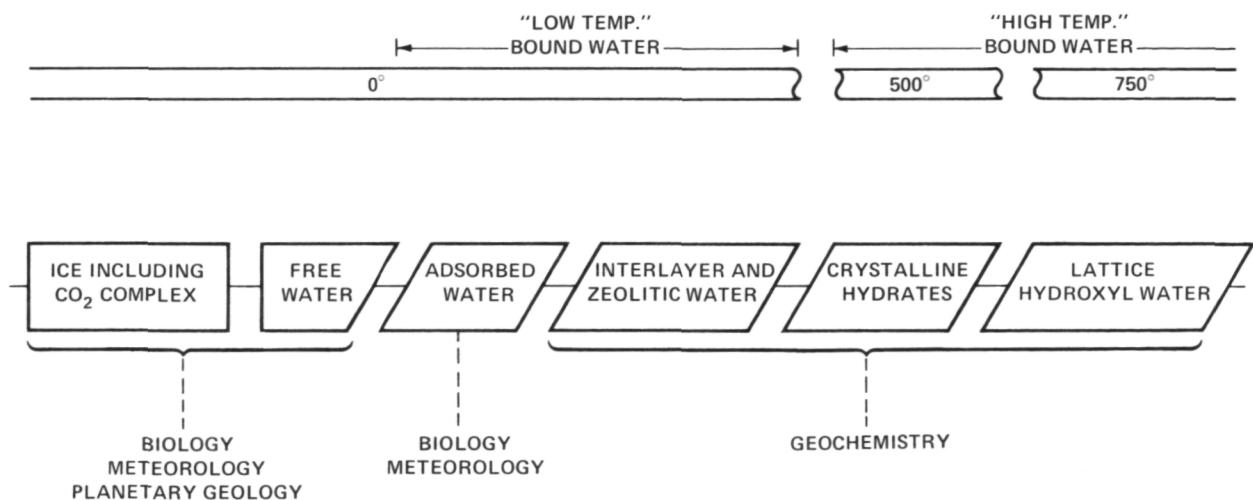


Figure 3.2. *Physical and chemical states of water expected on planetary bodies shown in relation to a temperature scale and in relation to geological, meteorological, and biological processes in which they are important.*

TABLE 3.2. EXPECTED FORMS OF REGOLITH WATER

| | Ices | Liquids | Adsorbed phases | Mineral hydrates |
|---------|------|---------|-----------------|------------------|
| Mercury | No | No | No | ? |
| Venus | No | ? | Yes | Yes |
| Earth | Yes | Yes | | |
| Mars | ? | ? | | |
| Jupiter | | No | | |
| Saturn | | | | |
| Uranus | | | | |
| Neptune | | | | |
| Pluto | | | | |

geochemical evolution of the body under investigation. Answers to these questions provide the information necessary to assess the availability of water for biological and geological processes.

If water is present on a planetary surface or in the regolith, it will nearly always be present in detectable quantities in the planetary atmosphere and vice versa. Detection of water in the atmosphere therefore answers the question of first priority. However, because no single general relationship between the quantities of atmospheric, surface, and subsurface water exists, the identification and measurement of soil water in its various forms at all locations of interest becomes an objective of high priority in all plan-

etary missions. This discussion considers the detection and measurement of water from landed spacecraft.

The general types of planetary water that landed spacecraft may be expected to encounter are illustrated in Figure 3.2 relative to a nominal temperature scale. This representation illustrates the ranges of temperature over which a given form of water is likely to exist. The diagram emphasizes the forms expected on Mars; however, if one bears in mind the fact that the lowest temperature forms include clathrate and other complex ice-containing mixtures, it represents the general aspects of the problem on other planetary bodies as well.

The forms in which water may appear in the regoliths of the principal planets are listed in Table 3.2, although some of the entries will be recognized as conjectural. For Venus and Jupiter, crucial data still being acquired bear directly on the question. Nevertheless, in spite of some of its controversial aspects, Table 3.2 conveys several important facts. Water vapor, although not shown explicitly, is the most ubiquitous. In the regolith, solid forms, including gas hydrate ices, and mineral hydrates ranging from adsorbed and nonstoichiometric associations through crystalline hydrates and secondary minerals containing "lattice" water, predominate. Apart from Earth, the liquid phase is rare and it is expected only in some of the venusian clouds and possibly as an ephemeral phase on Mars.

Gas hydrates are nonstoichiometric, clathrate structures composed of gaseous molecules frozen within cagelike configurations of water molecules. In discussing the possible occurrence of these substances throughout the solar system, Miller has pointed out that, although it is uncommon on Earth, an air hydrate appears to exist as a constituent of the Antarctic ice cap (Miller 1961; Miller 1969). Carbon-dioxide hydrate is a possibility on Mars (Milton 1974). Ammonia and mixed methane-hydrogen-benzene hydrates are possible on Jupiter, Saturn, and Uranus, and perhaps also on some outer-planet satellites.

The ideal planetary soil-water measuring instrument would (1) be specific for water, (2) be able to discriminate among the various possible forms of water, and (3) be capable of quantitative assay of both quantity and, if biological aspects are to be addressed, of availability. By international convention, the biological availability of soil water is discussed in thermodynamic terms. A potential function φ , called the *water potential*, has been defined for this purpose. It expresses, in terms of an energy difference, the availability of water held within a substance, supporting medium, or substrate relative to that of pure water in its natural liquid form or to that of vapor or ice. Figure 3.3 illustrates the various units and methods of mensura-

tion in common use. The ideal instrument would be relatively simple and direct in concept and implementation to minimize problems of compatibility with potentially available space vehicles. In considering deployment, the amounts of ice, frost, snow, liquid phases, adsorbed, and chemically combined forms of water should be measured as functions of location, depth, and time throughout the planetary regolith.

Typical of the most important geochemical questions associated with planetary missions are those that have been raised in connection with the Viking mission to Mars. For example, if the martian regolith has not reacted to an appreciable extent with the planetary atmosphere, the endogenic rock-forming processes will likely produce anhydrous silicate and oxide minerals. If, on the other hand, chemical weathering has occurred in the direction of equilibrating with a water-containing atmosphere, significant accumulations of hydrates, carbonates, and other volatile-containing minerals may have been formed (Anderson *et al.* 1972; Fanale *et al.* 1974). Similar mineral forms may have formed by chemical precipitation from a pre-existing hydrosphere. Detailed determinations of the water content of the materials forming planetary regoliths thus would yield valuable information bearing on rock type, degree of hydrothermal alteration and weathering, and position in the magmatic differentiation or the extrusive eruption cycles.

It was suggested some years ago that permafrost may be present on Mars (Otterman and Bonner 1966; Anderson *et al.* 1967; Wade and DeWys 1968; Briggs 1974). Permafrost is defined in terms of a temperature regime only, without reference to whether a soil or rock is ice-rich or dry; thus permafrost is soil that remains perpetually below 0° C. Since the mean temperature of Mars over the entire planetary disk is of the order of -50° C, it follows that permafrost surely exists (Öpik 1966). Critical questions (Leighton and Murray 1966; Anderson *et al.* 1972; Murray and Malin 1973; Sharp 1973) relate to the permafrost material: is the regolith ice-free or ice-rich? Do hygroscopic or

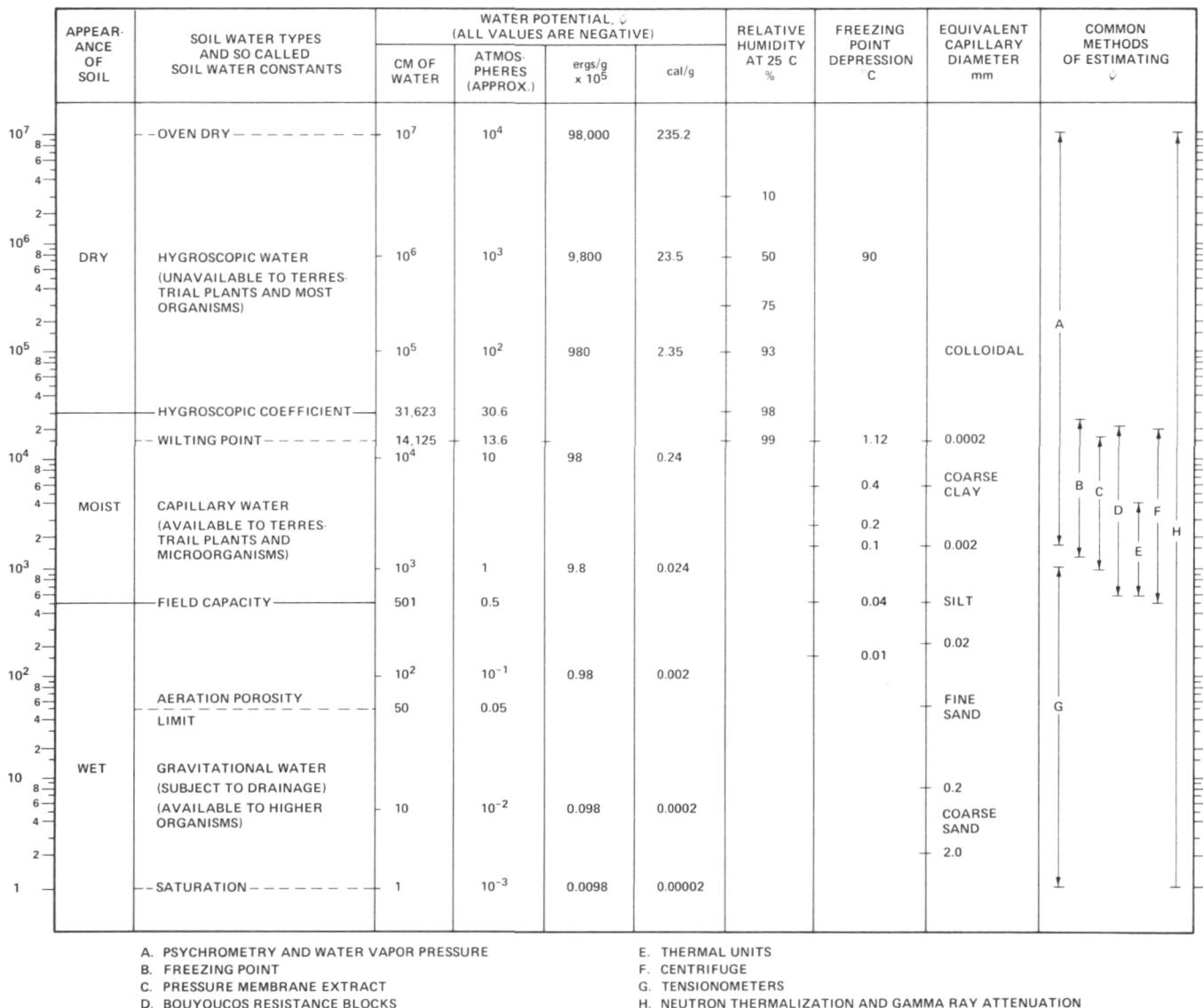


Figure 3.3 Common methods of measuring soil water.

deliquescent salts create unfrozen brines? To what degree does intergranular ice affect the bearing capacity and strength of the soils and rocks?

More than 200 established analytical methods for water analysis have been developed. Of the various analytical approaches, four are particularly promising: thermal evolution with an analysis of the evolved volatiles for water vapor, several spectroscopic methods, a radiochemical method, and several chemical methods involving solvent extraction and chromatography or titra-

tion. Methods that can be considered candidates for soft landers are (1) differential scanning calorimetry (DSC), in which the energy associated with any phase change in the sample is detected and measured, combined with an analysis of the evolved gases (EGA) for water; (2) solvent extraction coupled with a water analysis by chromatography; and (3) neutron moderation/gamma ray attenuation techniques.

The DSC-EGA method is well known in the laboratory and a science breadboard design adapted to soft landers has been developed. A

1- to 10-mg sample of surface material which has passed a 400- μm sieve is introduced into a small metal container. A vented lid is sealed to the top of the container, which is then placed into one of a pair of thermally balanced ovens. The ovens are covered and a purge gas is passed through the assembly. The ovens are heated at a constant rate of temperature rise and the differential power required to keep the two ovens at the same temperature is measured. This differential power is a measure of the energy of phase change occurring in the sample. During heating, any evolved gases are swept by the purge gas stream through a P_2O_5 electrolytic hygrometer which is specific for water and provides a quantitative measure of the amount evolved during the programmed temperature rise.

Solvent extraction and subsequent analysis by chromatography is also a well-established laboratory technique for water analysis. This method has been examined for compatibility with planetary vehicles and is considered feasible. Also promising are neutron moderation and scattering techniques. Science or engineering breadboard designs suitable for planetary soft landers based on these techniques, while considered feasible, have yet to be developed.

For hard landers, the choices seem limited at present to thermal evolution combined with effluent (evolved gas) analysis, by the P_2O_5 hygrometer, for example, or possibly by neutron moderation or neutron gamma-ray scattering techniques. The latter are possibilities only if suitable detectors can be hardened sufficiently for flight qualification. At present, these approaches are in the tentative, early stages of consideration. The principle of thermal evolution, followed by a specific detection and measurement of the water evolved on heating, is easily within reach. Adapted to a planetary penetrator, the instrument would consist of a suitably configured heater and a hardened P_2O_5 electrolytic hygrometer. Such a device is considered capable of accomplishing semiquantitative analyses of the regolith for water and ice as a function of longitudinal and latitudinal position and depth. It therefore closely approaches the criteria required to define the fundamental

aspects of planetary hydrology. An instrument of this type could easily detect ambient water-vapor contents at the part-per-million level and detect the presence of regolith water at levels of less than 1 percent by weight.

Summary

The measurement of regolith water and ice should be an important scientific consideration in all planetary, satellite, asteroidal, and cometary missions. In many instances, it will be important to address such questions as: the presence or absence of frozen, adsorbed, or chemically bound volatiles in the regolith; the characteristics of past surface environments and the mechanisms of weathering; the diurnal and seasonal exchanges of atmospheric and regolith water; and, in some instances, the availability of planetary water resources to living organisms. Three promising techniques have been identified to this end so far: DSC-EGA, solvent extraction/chromatography, and neutron, moderation/gamma-ray attenuation. One of these, the DSC-EGA technique, has been developed to the science breadboard stage for soft landers. A variant of this technique, thermal evolution with effluent gas analysis by electrolytic P_2O_5 hygrometer, is judged suitable for early planetary penetrator vehicles.

ORBITAL MEASUREMENTS

M. H. Carr

Many precise chemical mineralogical determinations are difficult to obtain from orbiting vehicles (Fig. 3.4). Nevertheless, extremely useful geochemical information can be gained from orbit. To evaluate the general implications of a few precise analyses from landers, we must have some knowledge of the geochemical variability of the planet's crust, and this is obtained practically only from orbital vehicles. By utilizing a variety of remote-sensing devices, geochemical variations can be mapped, geochemical anomalies identified, and limits placed on chemical and mineralogical compositions over wide areas. Unique chemical and mineralogical determina-

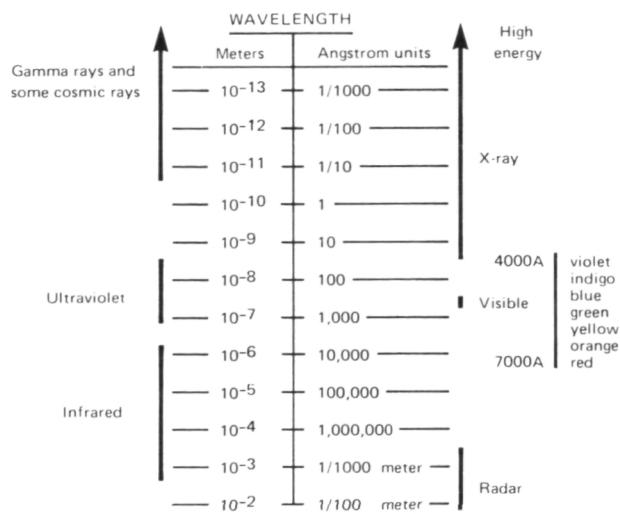


Figure 3.4. Electromagnetic spectrum showing the range of different spectral regions or working ranges for obtaining data by remote sensing.

tions cannot be expected from such measurements; in precision and generality, the results will be more like the determinations of terrestrial gravity and magnetic maps and thus subject to a limited set of interpretations.

Gamma-Ray Spectroscopy (A. Metzger)

A gamma-ray spectroscopy experiment to measure the composition of surface material to a depth of about 0.5 m can be undertaken from orbit around any planet or satellite body possessing little or no atmosphere. The gamma-ray spectrum coming from a planetary surface is composed of narrow discrete line spectra superimposed on a continuum. The line spectra are produced by gamma-ray emission associated with the decay of naturally radioactive elements (K, Th, and U), and from nuclear transitions in many other elements induced by the secondary neutrons derived from cosmic-ray bombardment. A polar or near-polar orbit will allow the entire surface to be mapped for those elements that provide a sufficient flux of characteristic gamma rays for detection. Besides the Moon, which has already been orbited by gamma-ray spectrometers, the technique is attractive for use at Mars (since gamma rays are only moderately

attenuated by the martian atmosphere), Mercury, those outer-planet satellites that are not immersed in a strong radiation belt or atmosphere, asteroids and cometary nuclei. At Mercury, the enhanced solar-particle flux during flares will provide an additional mechanism for the production of characteristic gamma rays that may be observed during and after the flare.

The energy and intensity of the line spectra carry the information needed to identify elements and to determine their concentration. The spectral resolution of the gamma-ray spectrometer is therefore important because this resolution defines the ability to separate spectral lines from each other and from the underlying continuum. The gamma-ray spectrometers that orbited the Moon as part of the science payloads on Apollos 15 and 16 were equipped with sodium iodide scintillation detectors with an approximate spectral resolution of 0.06 MeV at 1 MeV. Despite the relatively limited spectral resolution of this instrument, it was possible to detect and measure K, Th, Ti, Fe, Mg, and Si with useful accuracies (Metzger *et al.* 1974). Offering great potential for the future is the high-purity germanium detector that typically has a spectral resolution of about 0.002 MeV at 1 MeV, representing an improvement of a factor of 30 with respect to the scintillation detector. This translates into the expectation of measuring Al, Ca, U, O, H, C, and possibly Na, Mn, and S, along with improved sensitivity for those also detectable by a scintillation crystal.

Elemental distributions are directly related to the evolution of planetary bodies within the solar system. A knowledge of composition will reveal whether bulk accretion and magmatic differentiation have taken place in the formation and subsequent evolution of the body. If differentiation has been a factor, the regional distribution and abundance of certain of these elements will indicate its nature and extent. For the Earth and Moon, the composition of rock materials on or near the surface has been relevant to the average composition of the entire planetary body. There is therefore reason to suppose that remote measurements of the surface chemistry of other bodies will also be pertinent to their composition.

tions. Gamma-ray measurements of the relative abundance of volatile elements (potassium and sodium) and refractory elements (uranium, thorium, and titanium) will provide a test of the hypotheses of a volatile-rich Mars and a volatile-poor Mercury. Information on the near-surface presence and the regional distribution of significant abundances of water and solid carbon dioxide will be obtained by measuring concentrations of H, C, and O. For Mars, this could be a dynamic measurement in which the changes in polar-cap composition would be monitored during the martian year.

Germanium detectors must be cooled during operation and this may be done with a passive radiative system. The useful energy range of measurement is 0.2-10 MeV. Spectra would be accumulated simultaneously over this entire energy range. Sufficient coverage is required to provide the discrete characteristic features with adequate statistics for quality analyses. This amounts to about 1 hr of data for each resolution segment of area. The spatial resolution is on the order of the altitude above the planetary body for an isotropic detector; this can be improved by a factor of 5 if desired by the use of an active anticoincidence collimator.

X-Ray Spectroscopy (A. Metzger)

The interaction of solar x-rays with a planetary surface gives rise to characteristic x-ray emission. In this way, composition may be mapped for major elements whose K-shell transition lines are sufficiently low in energy to respond to the solar x-ray spectrum. Considering the shape of that spectrum, and the efficiency of gas proportional counters — which are currently the most suitable for such an experiment — Mg, Al, and Si are the elements most readily measurable. The successful Apollos 15 and 16 x-ray fluorescence experiments demonstrated the differentiated nature of the lunar highlands and pointed up the distinction between highland and mare regions by determining Al/Si and Mg/Si ratios over the entire portion of the ground track illuminated by the Sun (Adler *et al.* 1972a, b). Inferences regarding major rock types

can be derived from this information. The spatial resolution of the Apollo instrument was on the order of 50 km.

Application of this technique to other planetary bodies requires the absence of an absorbing atmosphere. Orbiting x-ray spectroscopy cannot be used at Mars; it can be used at Mercury and (based on what is known of their atmospheres) at any of the outer planet satellites with the exception of Titan. Asteroid missions are another potential application.

The observed solar flux varies in intensity according to distance from the Sun; consequently, the signal at Mercury will be greater than at the Moon. This can be taken advantage of in terms of greater sensitivity and improved spatial resolution. Conversely, experiments at greater distances from the Sun will see lower fluxes for a given amount of observing time; this will be a sizeable effect at Jovian distances and greater. A special opportunity for an orbiting or encounter x-ray experiment exists for the Galilean satellites that pass within the Jovian radiation belts. In this case, the exciting flux will consist of electrons. Their intensity and relatively high-energy spectrum should produce a measurable response from all abundant elements on the surface having an atomic number greater than oxygen or sodium, with the use of a silicon or germanium solid-state detector as the sensor. Good collimation against the background would be necessary. Alternatively, this experiment could be performed outside the radiation belt with a sensor of substantial detecting area ($10-100 \text{ cm}^2$) and a spacecraft pointing capability.

The Sun's x-ray flux varies over a time scale of minutes to days due to the creation of active regions on its surface. As the intensity increases over that of the quiet Sun, the spectrum generally contains a greater proportion of more energetic x-rays. Under these conditions, the prospects for detecting elements heavier than Si, such as Ca, K, and Fe, on Mercury, in addition to Mg, Al, and Si, will be favorable. The number and strength of these enhanced periods increases toward the maximum of the 11-yr solar cycle, which will next peak in 1979. Whereas the

Apollo experiment yielded Al/Si and Mg/Si ratios, results in terms of element concentrations should be possible with improved techniques for directly monitoring the solar x-ray spectrum and its temporal variations.

Far-Ultraviolet Radiometry (M. A. Steggert)

Far-ultraviolet radiometry is a technique, until recently neglected, which has the potential of providing some chemical information. Compositions can be inferred from values of the refractive index as derived by measurements of variations in the far-ultraviolet albedo. Measurements conducted with the Apollo 17 ultraviolet spectrometer (Fastie *et al.* 1973) have partially verified the applicability of this concept.

Briefly, the far-ultraviolet albedo of planetary surfaces is primarily determined by the refractive index of the material. As the majority of minerals are opaque in this spectral region, the far-ultraviolet albedo depends not on the volume reflectivity (hence body color) but rather on the surface reflectivity.

Lucke *et al.* (1973) noted that the lunar albedo increases with decreasing wavelength. This is to be expected because the imaginary part of the complex index of refraction of silicates also increases in this spectral region. In addition, the maria display a higher ultraviolet albedo than the lunar highlands; it is thus concluded that the mare material has a higher complex refractive index than the highland material.

A comparison of albedo ratios for two wavelengths would result in a determination of color differences. In principle, mineral species could be identified by comparing the refractive index and the spectral variations in the refractive index. Differences in refractive index could therefore provide information that would complement visible and near-infrared spectrophotometry.

One potential for the use of far-ultraviolet radiometry is its application to lava flows. Undegraded lava flows have a form and structure evident in photographs. However, if the flows occurred in the early stages of lunar surface evolution, subsequent processes such as meteor-

ite bombardment and day/night thermal processing will blur the structural forms, while the solar wind, which acts on the body color, will blur the optical color distinctions between the flow and its surroundings. The refractive index and hence the far-ultraviolet albedo should remain unchanged for longer periods, thus enabling the flow to be delineated.

Multiband Photometry in the Visible and Near-Infrared (J. Adams, T. McCord, and R. Huguenin)

Multiband photometry in the visible and near infrared is a technique for determining surface mineralogy and chemistry. The technique has already been applied with considerable success in studies of the Moon.

Reflection spectra are measured (0.3 to 2.5 μm) using Earth-based telescopes; the absorption bands present are interpreted from laboratory and theoretical studies of the optical properties of lunar, terrestrial, and meteoritic materials. The restriction of Earth-based telescopes limits the spatial resolution and wavelength range, mainly because of atmospheric turbulence and absorption. The most logical next step is the development of planetary orbiter spectroscopic experiments.

By placing a spectrometer in polar orbit around a planet or satellite, high-spatial-resolution (1–10 km) data of the spectral reflectance properties of the entire surface could be obtained, and these spectra would be free of the intense and variable terrestrial atmospheric absorption in the complete spectral region of interest (0.3 to 4.0 μm).

For the Moon, spectra from telescopic observations and laboratory studies of returned samples indicate that spectral reflectance measurements with 100–200 \AA resolution and 0.1 to 0.5 percent intensity precision over the 0.3 to 2.5 μm wavelength range would permit the determination of (1) regional variations in the average pyroxene, olivine, plagioclase, and glass compositions of rocks and (2) regional variations in the pyroxene/olivine, pyroxene/feldspar, and pyroxene/ilmenite abundance ratios. Techniques

have been developed that will permit the determination of the FeO and TiO₂ content of mature mare soils to about ± 1 percent. Soil maturity and relative exposure ages of surface material can be mapped. The spectra of fresh craters, which expose less altered material, contain information about the subsurface rocks.

Earth-based telescopic reflectance spectra, of 200- to 400-km-diameter regions on Mars, and laboratory data reveal that the bright surface material is composed in large part of oxidized weathering phases of fine-grained size, whereas the dark areas are relatively less weathered, coarser-grained, crustal material. The bright surface material is relatively uniform in composition across the surface, due probably to mixing by the intense global eolian activity. The dark area soils display regional variations in composition, which probably reflect regional variations in crustal composition. Clearly, high spatial resolution mapping of the surface would provide a great deal of information about regional and local variations in surface composition.

Ground-based and Mariner 6 near-infrared spectra of Mars reveal that reflectance spectra in the 1 to 4 μm region with an intensity precision of 0.1 to 0.5 percent and a spectral resolution of 100 to 200 \AA should be able to resolve H₂O vapor abundances of 3 to 5 precipitable microns or greater, and they should be able to separate water of hydration from water ice. The experiment should be able to detect ice or adsorbed H₂O layers greater than 0.1 μm thick and solid CO₂ layers greater than a few millimeters in thickness. Repeated spectral mapping of the polar regions could monitor seasonal variations in the condensed phase H₂O and CO₂ on the surface of the polar caps. Moreover, diurnal and seasonal variations in the amounts of H₂O condensed on the surface in the equatorial and mid-latitude regions could be monitored.

Ground-based reflectance spectra of Mercury indicate a Moon-like soil composed of iron-bearing and probably titanium-bearing glasses. Regional variations in optical properties and terrain, revealed by Mariner 10 images, probably correspond to regional variations in composition, as they do on the Moon. The use of

Earth-based telescopes permits only near-full-disk spectra. An orbiting reflectance spectrometer operating in the 0.3 to 2.0 μm region would obtain high spatial resolution data on the composition of these terrain units.

A spacecraft reflectance spectroscopy experiment also would be very useful in asteroid reconnaissance missions. From Earth-based telescopes, the spectra of over 100 asteroids indicate that almost all mineral assemblages found for asteroids exist among meteorites, although the frequency distribution of compositional types is different among the asteroids than among meteorites. Asteroids appear to be mostly homogeneous over their surfaces, displaying little variation in optical properties with rotational phase. Inner-belt asteroids appear to be more commonly of high-grade iron silicate composition, whereas outer-belt asteroids are more often of a primitive carbonaceous chondrite nature.

A spacecraft reflectance spectroscopy experiment would provide compositional information about the surfaces of the satellites of Jupiter and Saturn and Saturn's rings as well. From the Earth, only the four largest of Jupiter's satellites are bright enough to be studied easily and only full-disk spectra can be obtained. Europa, Ganymede, and Callisto all display absorption bands characteristic of H₂O ice in the 1 to 2.5 μm region, but the visible portion of the spectra contains strong, unidentified absorption, unlike that for ice.

Each of these applications could be carried out with a single instrument: a multi-dispersing-element spectrometer with a useful wavelength range of 0.3 to 4.0 μm , a spectral resolution of 50–200 \AA across the spectrum, and an intensity precision of 0.1 to 0.5 percent. Preliminary design indicates that this can be accomplished with an instrument that weighs just over 10 kg, consumes approximately 10 W of power, and has a bit transmission rate lower than 10^4 bits/sec⁻¹. Detectors would not need to be cooled and onboard calibrations would not be necessary, although they would be desirable. No developments beyond the present state of the art appear necessary.

Far-Infrared Spectroscopy (4–100 μm)

Spectroscopic studies of the emission characteristics of rocks and soils can provide diagnostic compositional information not available by any other remote-sensing technique. Silicates, for example, have characteristic emission minima in the 8–12 μm region (Reststrahlen features), which occur at the short-wavelength side for felsic rocks and at progressively longer wavelengths for mafic and ultramafic rocks. One limitation of the spectral studies has been the diminished spectral contrast that accompanies reduced grain size (Lyon 1964). Recent laboratory and theoretical studies hold some promise for extracting information from shifts in the emission maximum that occurs on the short-wavelength side of the strong Reststrahlen features. The position of this maximum is related to the Christiansen frequency (at which the index of refraction is unity and the extinction coefficient small) and is somewhat diagnostic of composition (Conel 1969; Logan and Hunt 1970).

Spectral studies of the Moon from Earth-based telescopes have been limited by the low-resolution, atmospheric absorption in diagnostic spectral regions and the fine-grained nature of the surface (Goetz 1968; Salisbury *et al.* 1970). Terrestrial studies of limited areas of good exposure have been more promising. Spectral emission curves have been used by Lyon and Patterson (1966) to discriminate various rock types by comparison with reference spectra. Because of the high data rates, this system is impractical for planetary mapping of an extended area. A mapping technique has been developed for spectral information that uses the ratio of adjacent bands measured by a dual-element detector (Vincent and Thompson 1972). The ratio, which is reasonably insensitive to moderate temperature variations, provides a measure of changes in position or shape of the emissivity minimum.

Bodies distant from the Sun receive less solar flux and consequently emit at lower temperatures, with maximum emission occurring at longer wavelengths. A consequence of this

reduced solar flux is that, for bodies of the same albedo, the crossover wavelength at which reflected and emitted energy are equal also shifts to longer wavelengths.

Useful application of the thermal inertia and spectral emissivity mapping techniques to planetary bodies beyond Jupiter is problematical. At such distances, the reduced thermal emission imposes severe instrumentation constraints that require a tradeoff between ground resolution and thermal resolution. Spectral emission measurements in the 10- μm region will be difficult not only because of the reduced emission but also because of the measurable contribution of reflected solar energy. Reststrahlen features occur in quartz in the 20–30 μm region and for other silicates out of 100 μm (Aronson *et al.* 1967). Thus diagnostic information exists at these longer wavelengths that can be used for geologic unit discrimination if the spectral features can be measured from orbit.

Recent laboratory studies have illustrated the importance of grain-sized distribution and temperature gradient on the appearance of emission spectra. Current theoretical models qualitatively predict some of the spectral features, but an adequate quantitative model has not yet been developed. Averaging spectral emission differences over heterogeneous terrain of varying compositions, surface temperatures, and temperature gradients is an imposing problem. The initial success of the ratio technique in terrestrial studies provides an example of the benefits gained when higher spatial resolution is obtained by sacrificing spectral resolution. This type of tradeoff may be necessary to obtain maximum geologic information for a given spacecraft data rate.

The spectral emissivity technique appears to be the most promising technique for compositional identification of materials from orbiters. But its current state of development and instrumentation constraints (e.g., higher data rates and lower signal/noise ratio for the same spatial resolution as the thermal mapper) generally make it a lower priority experiment. A lunar orbiter experiment could provide many of the answers as to the feasibility of the technique of

granular materials. If the results obtained are positive, the experiment would provide justification for a dedicated mission that the high data rates presumably dictate. Another alternative might be to use the image ratio technique. Again, good spectral data obtained from lunar orbit could be used to evaluate the feasibility of this technique for planetary exploration.

Comparison of the results from the Viking IRTM, which consists of several broad thermal bands, with the Mariner IRIS, which had good spectral resolution from 6 to 50 μm , will help determine the reliability of chemical identifications by comparatively broadband thermal radiometry.

Infrared Radiometry (H. Kieffer)

Measurement of the amount of thermal emission from a planet can provide information about the physical properties of the surface, the general chemistry of the surface, the composition of condensates, and indications of internal activity if large heat flows are found.

Broadband infrared measurements are usually expressed as brightness temperatures, which are a lower limit on the physical temperature of the object sensed. For geologic materials, whose emissivity is near unity, the brightness temperature is a good approximation of the surface temperature, and this information is directly useful in chemical or biologic theories.

Because infrared radiation penetrates less than a millimeter in geologic materials, the temperature measured represents a natural instantaneous balance between (1) the absorbed insolation, (2) emitted radiation, (3) surface heat flow, and (4) if there is an atmosphere, conductive and radiative phase transfer with the atmosphere and possibly latent heat of condensate phase transition. For the Moon, Mercury, and the Galilean satellites, only the first three are important and the direct analysis of brightness temperature in terms of thermal inertia, which governs the magnitude of the "daily" temperature variations, has been extremely useful (Pettit and Nicholson 1930; Murray 1967; Hansen 1973). The diurnal temperature variation, not readily sensed by any

other method, can be related to the surface particle size. The average value and surface distribution of particle size is very useful in determining the erosional history of a surface.

The sizes of many asteroids and satellites have been determined by measuring both their reflected and emitted energy (Matson 1971; Morrison 1973). Methodical observations of such telescopically unresolved objects will continue to improve our knowledge of the sizes of the smaller bodies in the solar system.

The subsurface heat diffusion is generally dominated by the diurnal term except near the poles, where the "annual" term is the largest. Heat flow from the interior is expected to be less than both of these terms except in local volcanic (in the general sense) areas. Thus the search by remote sensing for indications of current volcanic activity has an inherently small probability of success. Extensive, high-resolution coverage would be required to expect positive heat-flow indications even for the Earth (Kieffer *et al.* 1973). Thus far, no positive indication of heat flow from the interior has been found, although some lunar observations suggest either locally increased heat flow or unusually high thermal inertia (Hunt *et al.* 1968). The Galilean satellites have very small obliquities so that the poles received very little insolation. Their polar regions are probably the best place in the solar system to measure internal heat flow by infrared means.

There appear to be major differences in composition of the satellites of the major planets (Morrison and Cruikshank 1974; Kieffer and Smythe 1974). They are likely to have condensate surfaces, at least in their polar regions, and systematic infrared observations by spacecraft may represent the only way to determine the composition of the condensate and "rock" surfaces.

For rocks, the gross spectral emissivity determined by broadband infrared radiometry relates to the average chemical bond lengths and is therefore directly related to the bulk chemistry; while reflectance spectra are perhaps more readily obtained, many of these spectral features are electronic and can be strongly influenced by

minor components or by oxidation state.

For spacecraft observations, which avoid the attenuation of the Earth's atmosphere, broadband infrared measurements should be accompanied by broadband reflectance measurements so that the amount of heat diffusing from the surface material can be directly determined from a single measurement. Where the bolometric albedo is not well known beforehand, this combination represents a much more positive analytic tool for determining surface physical properties.

Infrared measurements of the Moon, combined with imaging and radar data of the same areas, have provided a much better geologic picture than can be obtained with these methods individually (Zisk *et al.* 1971). Measurements of this type are planned for Mars, where some radar coverage is available, and should be considered for the other accessible planetary surfaces.

Spacecraft instrumentation for infrared radiometry is already well developed. The main weight is in the collecting optics and the major design decision will be on the spatial resolution required. For exploratory missions, resolution of about 1/20 the object's diameter is desired so that latitude and diurnal variations can be observed and so that the temperatures of condensates covering only part of a body (as appears to be the case for the Galilean satellites) can be determined. Missions that plan to do geologic mapping, as do the martian orbiters, should have a resolution of about 5 to 20 km, which will allow discrimination of major geologic features, such as floors and rims of craters, tectonic scarps or rift valleys, volcanic constructs and flows.

Infrared radiometry has thus far been obtained from flybys and an orbital mission for Mars and the first flyby of Mercury. This has provided a single line of measurements or measurements that are widely spaced in the direction perpendicular to the orbital motion of the spacecraft. A large increase in the geologic interpretation of infrared radiometry is expected for systems that map approximately uniformly in two dimensions, as is planned for the Viking missions. This scanning capability should be

included for missions that anticipate resolution smaller than a tenth of the object's diameter.

Microwave Radiometry

Microwave radiometry is potentially a diagnostic remote-sensing technique. However, the complex interrelationships among the many parameters that control radiant energy have impeded fruitful application of the technique. Studies of thermal radio emission of the planets (Meyer 1970) and of the Moon (Weaver 1965), the quantitative determination of moisture content in soils (Poe *et al.* 1971), the determination of ice thickness (Adey *et al.* 1972), and the recognition of general potential for mapping layer thicknesses (Blinn *et al.* 1972) suggest that the situation may be changing.

In addition to its dependence on the four properties controlling radar reflections (see Ch. IV, "Radar Studies of Surface Properties") — dielectric constant, geometry, internal scattering, and absorption — microwave emission depends on the physical temperature and thermal gradient. A planet undergoes both diurnal and seasonal variations in surface temperature. A thermal wave communicates these temperature fluctuations to subsurface layers, but with depth the wave becomes progressively damped in amplitude and retarded in phase. Unlike radar, where most sensed energy comes from surface reflections, all of the thermal microwave energy originates at depth within the body. Because the overall microwave opacity of solids generally decreases toward longer wavelengths, microwave radiometers tuned to lower frequencies can observe at progressively greater depth. With increasing wavelength, we should therefore expect a decrease in the amplitude of the diurnal temperature oscillations and an increase in the phase lag.

The geologic interest in passive microwave measurements results from the possibility of determining internal heat flow and from determination of near-surface properties. Theoretically, the internal heat flow can be measured from the variation in mean surface temperature as measured at different wavelengths, provided

the thermal conductivities of the near-surface materials are known. In practice, surface properties rarely may be sufficiently well known to permit the unambiguous measurement of thermal gradients. Passive microwave measurements accurately predicted the lunar thermal flux of about $3 \mu\text{W}/\text{cm}^2$ as early as 1968.

The radar reflectivity varies with angle of incidence and polarization as well as with dielectric constant which is a function of rock type, moisture content, and in general, depth. The depth variation may introduce further view-angle variations. For a specular surface, power as a function of incidence angle may be used to obtain the dielectric constant. However, if knowledge of the brightness temperature is limited to a few view angles, or if the surface is nonspecular, then ambiguity arises between the determinations of thermal temperature and of dielectric constant.

The geometry of the surface (e.g., its roughness) strongly influences the absolute value, the polarization, and the angular dependence of the emissivity. Even if both polarizations are recorded as a function of incidence angle, it is difficult to distinguish between effects caused by roughness and those caused by dielectric constant for cases which are neither clearly specular nor clearly diffuse. In addition, the problem of interpretation would be compounded if interference caused by reflections between layers occurred.

Briefly, thermal microwave radiometry may, in particular instances, be used to infer any one of the following physical properties: physical temperature, temperature gradient, dielectric constant, or geometry (roughness and layering). The qualification for an appropriate application is that, for a quantitative result, five of the six parameters (temperature, temperature gradient, dielectric constant, geometry, absorption, and internal scattering) must be reasonably well known. They must be measured independently or assumed as part of a coherent physical model. This can rarely be achieved, but in view of the importance of internal heat flow, the development of microwave measuring techniques at several wavelengths and their possible

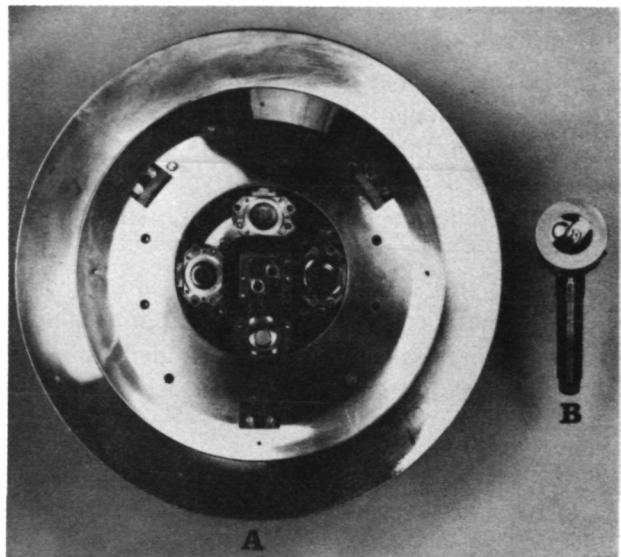


Figure 3.5. Comparison of the sensor heads of the lunar Surveyor alpha scattering instrument (A) and of the "mini-alpha" instrument (B). The alpha scattering instrument was flown on the Surveyor spacecraft in 1967-1968 and obtained the first chemical analysis of the lunar surface. The "mini-alpha" is an improved and miniaturized instrument developed for lander-type missions to obtain the chemical composition of extraterrestrial bodies. It operates in alpha, proton, and x-ray modes. The x-ray detector is cooled by a Joule-Thomson cryostat (gas supply not shown), which can provide an almost complete chemical analysis of a sample, with sensitivities down to the ppm range for many elements. (Courtesy of A. Turkevich)

applications to planetary problems should be supported.

SUMMARY

1. Geochemical work should begin with a determination of the mineralogical and petrological relations so that the processes by which major rock units formed can be inferred and the relative proportions of different rock types deduced.

2. Mineral assemblage is as important as bulk chemical composition because the mineralogy reflects both chemical composition and conditions of formation and is far more diagnostic of

rock type and formative processes than composition alone.

3. Difficult measurements such as isotopic and trace-element composition cannot be deferred with the expectation of early sample return to Earth, but must be performed remotely. The necessary hardware should be developed.

4. Since no single analytical technique can make all the necessary chemical and mineralogical determinations, emphasis should be placed on exploring the commonality of different systems so that an integrated chemical package can be assembled.

5. Orbital chemical measurements, such as infrared spectroscopy, although imprecise and subject to a variety of interpretations, are essential for interpolation between ground stations and for choosing the best places for ground measurements. Detailed chemical mapping from an orbiter should follow acquisition of surface images.

6. Concurrently with instrument development (Fig. 3.5), theoretical studies on stability relations under planetary conditions should continue, both to aid in instrument design and to ensure correct interpretation of the resulting data.

REFERENCES CITED

Adams, J. B., 1968. Lunar and martian surfaces: petrologic significance of absorption bands in the near infrared. *Science* 159:1453-1455.

Adey, A. W., Hartz, T. R., Barrington, R. E., Rolfe, W., and Mather, W. E., 1972. Theory and field tests of a UHF radiometer for determining sea ice thickness. Rept. of the Communications Research Center, Canadian Dept. of Commerce, Ottawa, Canada.

Adler, I., Trombka, J., Gerard, J., Lowman, P., Schmadebeck, R., Blodget, H., Eller, E., Yin, L., Lamothe, R., Gorenstein, P., and Bjorkholm, P., 1972a. Apollo 15 geochemical x-ray fluorescence experiment: preliminary report. *Science* 175:436-440.

Adler, I., Trombka, J., Gerard, J., Lowman, P., Schmadebeck, R., Blodget, H., Eller, E., Yin, L., Lamothe, R., Osswald, G., Gorenstein, P., Bjorkholm, P., Gursky, H., Harris, B., 1972b. Apollo 16 geochemical x-ray fluorescence experiment: preliminary report. *Science* 177:256-259.

Anderson, D. M., Biemann, K., Orgel, L. E., Oro, J., Owen, T., Shulman, G. P., Toulmin II, P., and Urey, H. C., 1972. Mass spectrometric analysis of organic compounds, water and volatile constituents in the atmosphere and surface of Mars: the Viking Mars lander. *Icarus* 16:111-138.

Anderson, D. M., Gaffney, E. S., and Low, P. F., 1967. Frost phenomena on Mars. *Science* 155:319-322.

Aronson, J. R., Emslie, A. G., Allen, R. V., and McLinden, H. G., 1967. Studies of the middle- and far-infrared spectra of mineral surfaces for application in remote compositional mapping of the Moon and planets. *J. Geophys. Res.* 72:687-703.

Blinn III, J. C., Conel, J. E., Quade, J. G., 1972. Microwave emission from geological materials: observations of interference effects. *J. Geophys. Res.* 77:4366-4378.

Brett, P. R., 1973. The lunar crust: a product of heterogeneous accretion on differentiation of a homogeneous Moon? *Geochim. Cosmochim. Acta* 37:2697-2703.

Briggs, G. A., 1974. The nature of the residual martian polar caps, *Icarus* 23:167-191.

Cameron, A. G. W., 1973. Accumulation processes in the primitive solar nebula. *Icarus* 18:407-450.

Chapman, C. R., Pollack, J. B., Sagan, C., 1969. An analysis of the Mariner 4 cratering statistics. *Astron. J.* 74:1039-1048.

Conel, J. E., 1969. Infrared emissivities of silicates: experimental results and a cloudy atmosphere model of spectral emission from condensed particulate mediums. *J. Geophys. Res.* 74:1614-1634.

Duncan, A. R., Erlank, A. J., Willis, J. P., and Ahrens, L. H., 1973. Composition and inter-relationships of some Apollo 16 samples. *Proc. Fourth Lunar Science Conf., Geochim. Cosmochim. Acta*, Suppl. 4, 2:1097-1113.

Fanale, F. P., 1971. A case for catastrophic early degassing of the Earth. *Chem. Geol.* 8:79-105.

Fanale, F. P., and Cannon, W. A., 1974. Exchange of adsorbed H_2O and CO_2 between the regolith and atmosphere of Mars caused by changes in surface insolation. *J. Geophys. Res.* 79:3397-3402.

Fanale, F. P., Johnson, T. V., and Matson, D. L., 1974. Io: A surface evaporite deposit? *Science* 186:922-925.

Fanale, F. P., Johnson, T. V., and Matson, D. L., 1975. Io's surface and the history of the Galilean Satellites. *Planetary Satellites, Proceedings, I.A.U. Colloquium, no. 28* (in press).

Fastie, W. G., et al., 1973. Ultraviolet spectrometer experiment: Apollo 17 preliminary science report. NASA SP-330, pp. 23-1 to 23-10.

Goetz, A. F. H., 1968. Differential infrared lunar emission spectroscopy. *J. Geophys. Res.* 73:1455-1466.

Grossman, L., and Larimer, J. W., 1974. Early chemical history of the Solar System. *Rev. Geophys. Space Phys.* 12:71-101.

Hansen, O. L., 1973. Ten-micron eclipse observations of Io, Europa and Ganymede. *Icarus* 18:237-246.

Hartman, W. K., 1973. Ancient lunar mega-regolith and subsurface structure. *Icarus* 18:634-636.

Houck, J. R., Pollack, J. B., Sagan, C., Schaack, D., Decker, J. A., Jr., 1973. High-altitude infrared spectroscopic evidence for bound water on Mars. *Icarus* 18:470-480.

Huguenin, R. L., 1974. The formation of goethite and hydrated clay minerals on Mars. *J. Geophys. Res.* 79:3895-3905.

Hunt, G. R., Logan, L. M., and Salisbury, J. W., 1973. Mars: Components of infrared spectra and the composition of the dust cloud. *Icarus* 18:459-469.

Hunt, G. R., Salisbury, J. W., and Vincent, R. K., 1968. Lunar eclipse: infrared images and an anomaly of possible internal origin. *Science* 162:252-254.

Johnson, T. V. and Fanale, F. P., 1973. Optical properties of carbonaceous chondrites and their relationship to asteroids. *J. Geophys. Res.* 78:8507-8518.

Kieffer, H. H., Chase, S. C., Jr., Miner, E., Munch, G., and Neugebauer, G., 1973. Preliminary report on infrared radiometric measurements from the Mariner 9 Spacecraft. *J. Geophys. Res.* 78:4291-4312.

Kieffer, H. H., and Smythe, W. D., 1974. Frost spectra: comparison with Jupiter's satellites. *Icarus* 21:506-512.

Larson, H. P. and Fink, U., 1975. Infrared spectral observations of asteroid (4) Vesta. *B.A.A.S. (in press)*.

Leighton, R. B., and Murray, B. C., 1966. Behavior of carbon dioxide and other volatiles on Mars. *Science* 153:136-144.

Levine, J. S., and Riegler, G. R., 1974. Argon in the martian atmosphere. *Geophys. Res. Lett.* 1:285-287.

Lewis, J. S., 1971. Satellites of the outer planets: their physical and chemical nature. *Icarus* 15:174-185.

Lewis, J. S., 1972. Metal/silicate/fractionation in the solar system. *Earth Planet. Sci. Lett.* 15:286-290.

Lewis, J. S., 1974. The temperature gradient in the solar nebula. *Science* 186:440-443.

Logan, L. M. and Hunt, G. R., 1970. Emission spectra of particulate silicates under simulated lunar conditions. *J. Geophys. Res.* 75:6539-6548.

Lucke, R. L., Henry, R. C., and Fastie, W. G., 1973. Far ultraviolet reflectivity of lunar dust samples: Apollo 11, 12, and 14. *Astron. J.* 78:263-267.

Lyon, R. J. P., and Patterson, J. W., 1966. Infrared spectral signatures — A field geological tool. *Proc. 4th Symposium on Remote Sensing of Environment, Univ. of Michigan*, pp. 215-230.

Matson, D. L., 1971. Infrared observations of asteroids. In T. Gehrels, ed., *Physical studies of minor planets*, NASA SP-267, pp. 45-50.

McCord, T. B., Adams, J. B., and Johnson, T. V., 1970. Asteroid Vesta: Spectral reflectivity and compositional implications. *Science* 168:1445-1447.

McCord, T. B., Elias, J. H., and Westphal, J. A., 1971. Mars: The spectral albedo (0.3-2.5 μ) of small bright and dark regions. *Icarus* 14:245-251.

Metzger, A. E., Trombka, J. I., Reedy, R. C., and Arnold, J. R., 1974. Element concentrations from lunar orbital gamma-ray measurements. Proc. Fifth Lunar Science Conf., *Geochim. Cosmochim. Acta*, Suppl. 5, 2:1067-78.

Meyer, C. H., 1970. Thermal radio emission of the planets and Moon. In Dollfus, A., ed., *Surfaces and Interiors of Planets and Satellites*, New York, Academic Press.

Miller, S. L., 1961. The occurrence of gas hydrates in the solar system. *Proc. Nat. Acad. Sci.* 47:1798-1808.

Miller, S. L., 1969. Clathrate hydrates of air in Antarctic ice. *Science* 165:489-490.

Milton, D. W., 1974. Carbon dioxide hydrate and floods on Mars. *Science* 183:654-656.

Morrison, D., 1973. Determination of radii of satellites and asteroids from radiometry and photometry. *Icarus* 19:1-14.

Morrison, D., and Cruikshank, D. P., 1974. Physical properties of the natural satellites. *Space Sci. Rev.* 15:641-739.

Murray, B. C., 1967. Infrared radiation from the daytime and nighttime surfaces of Mercury (abstract). *Trans. Am. Geophys. Union* 48:148-149.

Murray, B. C., and Malin, M. C., 1973. Polar volatiles on Mars — theory versus observation. *Science* 182:437-443.

O'Kelley, G. D., Eldridge, J. S., Schonfeld, E., and Northcutt, K. J., 1972. Primordial radio-elements and cosmogenic radionuclides in lunar samples, from Apollo 15. *Science* 175:440-442.

Öpik, E. J., 1966. The martian surface. *Science* 153:255-265.

Otterman, J., and Bronner, F. E., 1966. Martian wave of darkening: A frost phenomenon? *Science* 153:56-60.

Owen, T., and Mason, H. P., 1969. Mars: Water vapor in its atmosphere. *Science* 165:893-895.

Pettit, E., and Nicholson, S. B., 1930. Lunar radiation and temperatures. *Astrophys. J.* 71:102-135.

Philpotts, J. A., Schnetzler, C. C., Nava, D. F., Bottino, M. L., Fullazar, P. R., Thomas, H. H., Schuhmann, S., and Kouns, C. W., 1972. Apollo 14: some chemical aspects. Proc. Third Lunar Science Conf., *Geochim. Cosmochim. Acta*, Suppl. 3, 2:1292-1305.

Pilcher, C. B., Ridgway, S. T., and McCord, T. B., 1972. Galilean satellites: Identification of water frost. *Science* 178:1087-1089.

Poe, G., Stogryn, A., and Edgerton, A. T., 1971. Determination of soil moisture content using microwave radiometry. Final Tech. Rept. 1684R-1, Microwave Div., Aerojet-General Corp.

Rubey, W. W., 1951. Geological history of sea water: an attempt to state the problem. *Bull. Geol. Soc. Am.* 62:1111-1148.

Salisbury, J. W., Vincent, R. K., Logan, L. M., and Hunt, G. R., 1970. Infrared emissivity of lunar surface features: 2. Interpretation. *J. Geophys. Res.* 75:2671-2682.

Schorn, R. A., Farmer, C. B., and Little, S. J., 1969. High dispersion spectroscopic studies of Mars: III. Preliminary results of 1968-1969 water vapor studies. *Icarus* 11:283-288.

Sharp, R. P., 1973. Mars: South polar pits and etched terrain. *J. Geophys. Res.* 78:4222-4230.

Short, N. M., 1975. *Planetary Geology*. Prentice-Hall, New Jersey.

Sokolov, S. S., 1974. Soviet automatic interplanetary stations investigate Mars. *Akad. Nauk S.S.R. Vestnik* 10:21-38.

Toulmin III, Priestley, Baird, A. K., Clark, B. C., Klaus, K., and Rose, H. J., Jr., 1973. Inorganic chemical investigation by x-ray fluorescence analysis: The Viking Mars lander. *Icarus* 20:153-178.

Turkevich, A. L., 1973. The chemical composition of the lunar surface. *Accounts Chem. Res.* 6:81-89.

Veeder, G. L., Johnson, T. V., and Matson, D. L., 1975. Narrowband spectrophotometry of Vesta. B.A.A.S. (in press).

Vincent, R. K., and Thompson, F. J., 1972. Rock-type discrimination from ratioed infrared scanner images of Pisgah Crater, California, 1972. *Science* 175:986-988.

Wade, F. A., and DeWys, J. N., 1968. Permafrost features on the martian surface. *Icarus* 9:175-185.

Weaver, H., 1965. The interpretation of thermal emission from the Moon. In *Solar System Radio Astronomy*, Plenum Press.

Zisk, S. H., Carr, M. H., Masursky, H., Shorthill, R. W., and Thompson, T. W., 1971. Lunar Apennine-Hadley Region: Geological implications of earth-based radar and infrared measurements. *Science* 173:808-812.

CHAPTER IV. GEOPHYSICS

M. H. Carr and P. Cassen

Geophysical investigations of the planets seek solutions to some of the most fundamental problems of planetology (Fig. 4.1). Necessarily interrelated, these problems include determination of a planet's structure and bulk composi-

tion, the present physical state of its interior and the nature of the processes that have shaped it, the degree and timing of differentiation, and, ultimately, by synthesis of aspects of all of the above, the thermal evolution of the planet.

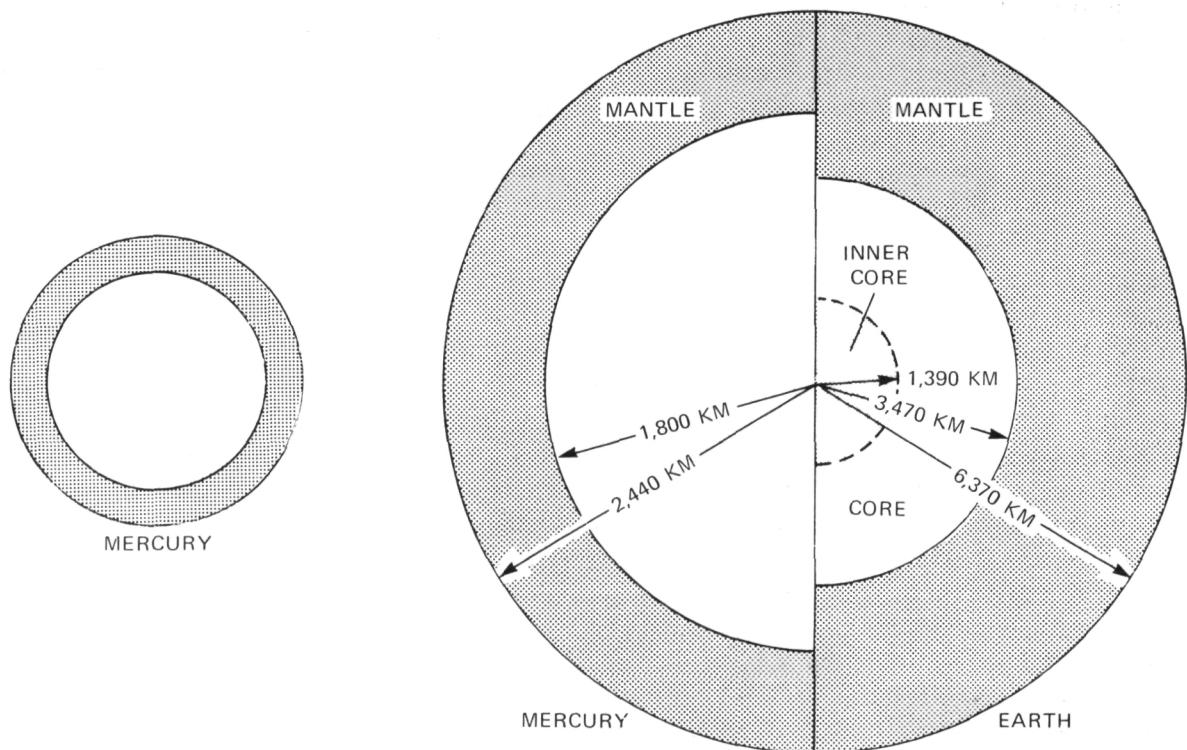


Figure 4.1. Cross-sectional views of Mercury and Earth scaled to the same diameter to show proportionately how much larger Mercury's iron core is thought to be than Earth's core. Disk at the left is Mercury drawn at the same scale as Earth. Knowledge of planetary interiors is derived from data on the figure, mass, geochemical properties, etc., and is fundamental in understanding the origin and evolution of the solar system (after Murray 1975).

Some of these questions can only be answered by *in situ* measurements. However, many geophysical properties, such as size, shape, dynamical figure, gravity field, global magnetic characteristics, and some surface properties can be determined remotely, either from spacecraft or from the Earth. Mapping of regional provinces by reconnaissance photogeologic methods and the determination of global topography, gravity, and magnetism provide a basic framework from which exploration can proceed. Later *in situ* measurements of seismicity, heat flow and thermal properties, local magnetism, and composition can then provide essential information that cannot be obtained by remote means.

Four areas of investigation, each dealing with the measurement of a particular geophysical property, are discussed in the following sections. These properties are the gravity field, seismicity, magnetism, and heat flow (see Kaula, 1968, for a general discussion). All are strongly affected by conditions, past or present, in the planetary interior; their measurement is our primary source of information about planetary interiors.

GRAVITY

S. Saunders

Because the external gravity field of a planet is determined solely by the distribution of mass within it, knowledge of the field puts strong constraints on the internal structure of the planet and provides data for models of composition and degree of differentiation. Compositional models must reproduce the measured mean density and moments of inertia. The moments of inertia indicate the degree to which mass is concentrated toward the center. If the planet is rotating rapidly (like the Earth) and is in hydrostatic equilibrium, the polar moment is given by the Radau-Darwin formula. If the precession of the axis of rotation caused by the Sun or other planets can be measured, the absolute values of the moments of inertia can be determined without recourse to the Radau-Darwin approximation. Deviations of the actual moment from this formula indicate departure from the hydrostatic state and may be related to the

degree of mechanical support or dynamic processes that exist in the interior. However, precessions are not presently known for planets other than the Earth and Moon.

The gravitational attraction of a planet can be ascribed to the sum of the attraction of a reference spheroid plus an anomalous component. The anomalous component can be related to effects of surface topography plus a contribution from internal mass variations. The *Bouguer correction* is the removal of topographic effects and the resultant Bouguer gravity is the anomalous gravity due to internal density variations only.

The Bouguer gravity distribution carries a number of geophysical implications. First, lateral variations are highly indicative of planetary differentiation. Correlation of the Bouguer gravity with topography will lead to an estimate of the degree of isostatic compensation — the tendency for the internal mass to rearrange itself to balance topographic loads. If a topographic feature can be dated by crater statistics, then the age of a feature combined with the Bouguer gravity can lead to an estimate of the time-averaged viscosity of the shallow interior. The viscosity, in turn, can be related to temperature and it may thus be possible to place constraints on thermal evolution. Comparison of Bouguer gravity with topography may also suggest the action of other physical processes such as plate tectonics and interior convection.

The radial distribution of density that satisfies the Bouguer gravity is, of course, not unique, but useful constraints may be applied. For example, interior density models may be constructed that minimize shear-strain energy. If the planetary crust is in a state of isostasy, and the Bouguer gravity is dominantly due to lateral variations in crustal thickness, then the absolute crustal isopach may be mapped. With or without an isostatic assumption, the combination of seismological constraints and gravity data is very powerful. With the Bouguer gravity (but not isostatic) assumption stated above, crustal thickness determined at a single point seismically will allow mapping of the crustal isopach. With the crustal isopach mapped independently by

seismic techniques, then with neither of the assumptions, the crustal contribution to the global Bouguer gravity may be removed. In turn, deep density variations may then be mapped, which might possibly shed light on inhomogeneous accretion or deep convection.

Gravity data may be obtained by observing the motions of spacecraft as they fly by or orbit a planet. Planetary flybys are limited in the resolution they can provide since they are in the vicinity of the planet for such a short time — about 1 hr for a small planet and perhaps 10 hr for the giant planets. The second-degree terms in the spherical harmonic expansion of the gravity potential is about the limit for flybys of the terrestrial planets. For the giant planets, the zonal spherical harmonics up to J_6 can be determined. Orbiting spacecraft can yield much better resolution of the gravity field because they provide extensive surface coverage, so local mass anomalies are readily detected and easily associated with particular topographic locations. Present technology is such that polar orbiters of Mars and Mercury at 100-km altitudes can provide complete detailed gravity maps with global resolution better than that for the Earth. These mappings would furnish first-order data for comparative planetology.

The limiting resolution of the gravity-field measurement is approximately equal to the altitude of the spacecraft. The best orbit is a close, circular, polar orbit; if this is impossible, several different orbits with various inclinations would be desirable.

The actual gravity determinations are made by observing the Doppler shift of a radio signal transmitted from the spacecraft relative to the receiver on Earth. In practice, a signal of known frequency is sent from the Earth to the spacecraft where it is given a precise frequency shift and relayed back by a transponder. The effects of primary orbital motion, Earth rotation, atmospheric drag, spacecraft outgassing, and the gravitational perturbations of other planets, the Sun, and satellites must be modeled well enough that they may be subtracted from the data. The global field can then be modeled with spherical harmonic fits, or the local field

may be determined by inversion of Doppler residual data. Gravity observations taken in this manner are routinely reduced for orbit determination during mission operations; that is, the gravity experiment is a refinement of procedures that must be done in the course of mission navigation.

There are other methods for determining the gravity field of a planet. Orbits of the natural satellites may be analyzed in the same way as those of artificial satellites, but, because of limitations in the accuracy of observations and the generally unfavorable (low) inclinations of the orbits, information obtained in this way is severely restricted. Information on the shape of the gravity field of Mars was obtained, or can be derived, from the spectral experiments involving atmospheric CO₂ absorption of light performed from Earth and onboard the Mariner 9 spacecraft. It has been proposed that analogous measurements can be made of the venusian field at microwave frequencies using Earth-based delay — Doppler interferometry.

Global gravity-field determinations of the planets are available for the Earth, Moon, and Mars. Local resolution is extremely good for the Earth because of widespread surface observations; however, since the oceans and polar regions do not have dense samplings, the global field needs refining. The resolution for the near-side of the Moon within the 30° north and south latitude band varies from 50 km to 200 km, and for Mars is about 1000 km in the latitude band near the Mariner 9 periapsis, about 23° S.

Venus, Mercury, and Jupiter have been observed from flyby missions. The Mariner 10 flyby of Venus and Mercury yielded data up through the second-degree terms in the spherical harmonic expansion of the gravitational potentials. These are of great interest with regard to questions of the spin stability and spin-orbit coupling of Venus. The Pioneer 10 and 11 celestial mechanics experiments have obtained good values of J_2 , J_4 , and J_6 for Jupiter. As expected, the observed values of J_3 and J_5 are approximately zero (if the planet is in hydrostatic equilibrium).

Gravity data are usually presented on planet-wide charts (typically 1:25-million scale) on which lines of equipotential are drawn. In this way, correlations with topography can be observed and interpreted (see Ch. V, Geodesy and Cartography).

MAGNETOMETRY

Planetary magnetic fields may be divided into three types: internal dynamo fields, induced fields, and remanent fields. The Earth's magnetic environment is dominated by an internal dynamo field, and it appears that at least two other planets, Jupiter and Mercury, possess such fields. Recent reports of synchrotron radiation from Saturn suggests a dynamo for that planet also. Although the theory of dynamos is far from complete, the minimum requirement for a planetary dynamo certainly can be taken as motion within a fluid region of high electrical conductivity. Thus, for a terrestrial planet at least, the existence of a dynamo is a strong indication of a molten metallic core, presumably composed mainly of iron since this is the most abundant metal. Whether the motion sustaining the dynamo is due to coupling of the fluid core (or layer) to the mantle, and thus to the planet's overall motion, or due to the release of thermal energy within the core is not yet clear. It is widely held that rotation of the planet is necessary. But the recently discovered field of Mercury indicates that the spin requirements may be minimal since Mercury rotates very slowly. A thermal source of energy would have far-reaching implications for planetary evolution. Whether the thermal source is heat of fusion from a solidifying core or the decay of radioactive elements, transport of this energy through the planetary mantle presents problems that have not yet been resolved for the Earth. For instance, a core that is thermally convecting will provide a heat flux to the mantle equal to at least the product of the thermal conductivity of the core and the adiabatic temperature gradient of the core at the core-mantle boundary. For the Earth, this corresponds to a considerable fraction of the heat flow from the surface; the impli-

cation is that much of the Earth's heat sources may be in the core despite the fact that the crust is enriched in radioactive elements. Clearly, comparative studies of other planets with dynamo fields are desirable.

An ordered, dipolar field within a solar-wind-induced bow shock and magnetopause is characteristic of a magnetosphere caused by an internal dynamo. Such a field can be detected by a spacecraft-borne magnetometer directly, as for Mercury, or by reception of synchrotron radiation originating in the magnetosphere, as for Jupiter and possibly Saturn. In either case, a magnetometer onboard a spacecraft with an eccentric orbit is required to map a magnetosphere. For planets with significant induction fields such as Venus (see below), this may be the only way to distinguish intrinsic dynamo fields from those associated with solar-wind atmosphere interactions.

Induced fields result from electric currents generated within a planet or its ionosphere as the changing magnetic field in the solar wind sweeps past. They may also occur in satellites of planets with dynamos as the satellite moves within the planetary magnetosphere. Measurement of the induced field permits inferences to be made about the current-carrying region; if it is within the planet, the electrical conductivity of the planetary material will affect the response induced by fluctuations in the solar-wind (or other external) field. Simultaneous measurements of the driving and induced fields then yield information on the conductivity-depth relation, which, in turn, can restrict the thermal profile of the interior of the planet. This experiment has been performed for the Moon (Dyal *et al.* 1974), where the driving field was measured by a magnetometer on an orbiting spacecraft, while the total field, including the induced component, was measured by magnetometers placed at the Apollo landing sites. The practicality of this experiment depends to a large degree on the magnetic and plasma environment of the planetary body. The interaction of the Moon with the solar wind is substantially simpler than that of most planets since the Moon possesses neither a significant atmosphere nor dynamo field. Thus

the driving field is easily identified with the solar-wind field, which, to a certain approximation, may be taken as homogeneous over the entire Moon. This is not the case for Venus, for instance, where there is a direct interaction of the solar wind with the atmosphere, and induced currents exterior to the planet are expected to be strong, local, and variable. Similarly, Earth-based studies have required the accumulation of magnetic field data spanning many years in order to calculate the mantle electrical conductivity. Further studies of the magnetic and plasma environments of the terrestrial planets are necessary to assess the possibility of using induced fields as probes of the interiors of these planets. Low-frequency or steady-state induction can also be used to measure the magnetic permeability and iron content of the outer regions of planetary bodies similar to the Moon (Dyal *et al.* 1974).

Measurements of remanent magnetism can, in principle, provide clues to the intensity and nature of ancient magnetizing fields and the temperature regime experienced by the magnetized specimen. The Moon exhibits considerable remanent magnetism (Fuller 1974); in fact, the source of the magnetizing field is one of the outstanding problems in lunar study. If an internal dynamo is required, severe constraints on thermal evolution are implied, as mentioned above.

Detailed analyses of returned samples would be necessary to extract the maximum paleomagnetic information for a planet because only in this way could the process and epoch of magnetization be directly established. However, detection of remanent fields can be accomplished by spacecraft, as has been done for the Moon. First, they may be measured directly by surface magnetometers or by low-orbiting spacecraft-borne instruments. Secondly, the effects of regions of relatively strong magnetization may be detected from an orbiting spacecraft if they interact with the solar wind. It has been found, for instance, that regions of enhanced remanence on the Moon produce shock waves in the solar wind when they appear at the limb. Finally, identifiable components of

the solar electron flux may be reflected by remanent fields, as happens for the Moon (Lin *et al.* 1976). An orbiting spacecraft with a plasma detector can then map regions of "electron reflectivity" and effectively determine areas of remanent magnetization. All of these methods rely heavily on the absence of strong induced and dynamo fields and are therefore best suited to Moon-like objects, such as asteroids, and perhaps the martian satellites.

The outer planets present important and, in some cases, unique problems for magnetometry, but the detection and mapping of their magnetospheres are prerequisites for inferring properties of their interiors. It is likely that techniques developed for the magnetic exploration of the terrestrial planets will also be useful for the exploration of the outer-planet satellites.

SEISMOLOGY

A seismology experiment provides information on three basic properties of a planet: its seismicity, internal structure, and composition (Fig. 4.2). Seismicity concerns the level of

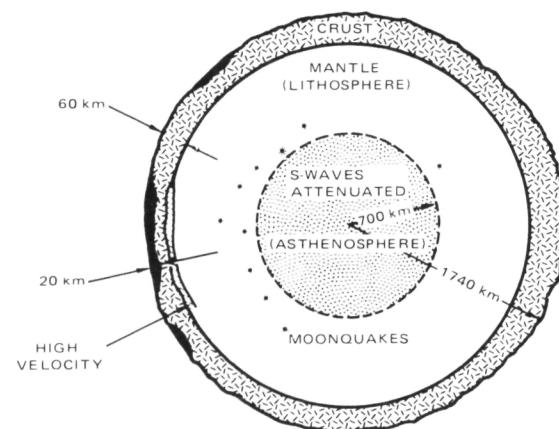


Figure 4.2. Cross section of the Moon showing the crust (thickness exaggerated), lunar mare areas (black), lithosphere, and asthenosphere. Moonquake foci are shown with stars. Data are derived primarily from seismometer stations established during Apollo missions (from Dainty *et al.* 1974).

seismic activity, the location of events, their correlation with geologic features, the nature of the stresses acting, and so forth. Knowledge of a planet's seismicity allows assessment of the dynamic state of the planet, places limits on thermal conditions of the interior, and constrains models of the evolution of the planet. Seismology is the principal means of determining internal structure — where discontinuities exist, whether the planet has a core, and how thick the crust is. The absolute values of seismic velocities also provide the most direct evidence on internal compositions and physical conditions. A seismic experiment should accordingly be given a very high priority in any exploration program.

The ideal seismic experiment requires a global network of seismic stations, each with a wide variety of seismic detectors and the necessary power and facilities for data handling. The possibility of establishing such a network on any planet other than the Earth is exceedingly remote, but this does not rule out a meaningful scientific experiment. Much can be learned from a relatively simple seismic experiment which would not only provide basic scientific data but enable better subsequent design of more complex and comprehensive systems.

The simplest experiment involves placing a single seismic detector on the surface of the planet, preferably a short-period (1 sec) single-axis-type seismometer; or, if telemetry and weight restraints permit, a three-axis type. From such a detector, the seismicity or aseismicity of the planet could be established and estimates made of the magnitude and frequency of seismic events. The general level of seismic noise could be measured, and from the separation of the S and P waves, approximate distances of events from the station could be inferred. In addition to its intrinsic scientific worth, the data would permit an accurate assessment of the advisability of more complex seismic stations for future missions and provide design constraints on such systems. A simple seismometer should be deployed at the earliest possible opportunity, preferably on the first lander to any planet. Other considerations, however, may dictate

deployment by later missions.

A second-generation seismic station should include both long- and short-period three-component seismometers. From such a station, a far more complete picture of the seismic properties of the planet can be obtained than from records of a single short-period instrument. Comparison of the amplitude and separation of surface and body waves enables distances and focal depths to be estimated. Distances and directions can be established with far greater accuracy than with a single short-period instrument, and better determinations of attenuation rates are possible. In addition, the seismic properties of the crustal materials can be deduced from the dispersion of surface waves resulting from the wave-guide effect.

A broad-band or long-period instrument can, in addition, give information on free oscillation periods and density. In principle, a single broad-band instrument, with response out to about 20 min for Mars and Mercury and 60 min for Venus, could provide detailed information on the structure of the planet in the spherically symmetric first approximation if a single large event were to occur during its operation. These events are rare on Earth (one every 3 or 4 yr), but moderate-sized events can generate long-period surface waves that can be used to determine structure and composition of outer layers (crust, upper mantle, lithosphere). Because of the large stresses on Mars (from considerations of its shape and gravity field), it may be a very active planet and, in fact, have large Marsquakes.

The effectiveness of a seismic station is vastly enhanced by the simultaneous operation of a duplicate station. Many of the ambiguities of interpretation are removed by having duplicate records, and seismic events can be more accurately located by using widely spaced stations. Far better estimates can be made of the variation of seismic velocity with depth than from a single station and, under certain conditions, shadow zones can be detected and the presence of a core determined. Consideration therefore should be given to the simultaneous operation of two distantly spaced stations early in the exploration program. Gravity-, tilt-, and

strain-variometer experiments to examine free oscillations, tidal effects, and secular deformation should be deferred until at least two stations have been established on the planet.

The use of penetrators, if feasible, would be an attractive way to install a seismic array on a planet. In addition to multiple coverage, essential for meaningful seismic experiments, the subsurface environment of the planet can considerably increase the signal/noise ratio and sensitivity of the instrument. The Viking experiment, for example, is limited — not only because two instruments will be emplaced, but also because the lander resonances decrease the sensitivity. In addition, wind noises are a problem on Mars, particularly if the lander has resonances in the seismic band, which seem unavoidable except in the case of penetrators.

Deployment of two three-axis, short-period seismometers is already planned for Mars. These should be followed by several six-component seismometers and long-period instruments, possibly deployed by penetrators. Particular emphasis should be given to long life.

The internal constitution of Mercury is of special interest because of the planet's high density. Although no landers are planned at present, a lander is feasible, and a simple seismic experiment should have high priority on the first lander.

A thorough comparison of Venus and Earth is necessary for an understanding of what factors control planetary evolution. A primary objective is to determine whether Venus, like the Earth, is divided into core, mantle, and crust. This is a problem that can be solved by seismic experiments, and the feasibility of designing a seismic experiment to survive in the $\sim 700^{\circ}\text{K}$ and 100 atm pressure that prevail on Venus should be explored.

Development of lightweight, compact, low-noise seismometers is well advanced. Instruments have been designed that weigh less than ~ 2 kg and that can withstand large loadings. The primary advance in recent years has been in electronics. It is now possible to design a lightweight seismometer with its own computer so that the pass band, gain, trigger thresholds, etc., can be

modified after landing to take advantage of information gained while on the surface. The principal problem in establishing a seismic station is not in detecting and recording the signal but in transmitting the information. Most of the seismic experiments on the Moon were based on technology of data analysis derived from our Earth-based experience. To obtain relatively few numbers, we analyze thousands of data bits from a seismogram. There are many possible ways to reduce the quantity of information required for transmission such as event-triggered, and these ways should be explored.

A seismic station must of necessity operate for a considerable length of time before a statistically significant number of events is recorded. No estimate can be made of this time until the level of seismicity of each individual planet is obtained. A minimum 6-month operation time should be aimed for, and the appropriate power sources for such lengthy operation should be developed.

Active seismic experiments are normally used to study near-surface discontinuities and near-surface seismic velocities. These are considerably less important than analysis of natural seismic events, and this type of active seismic experiment is given low priority on a seismically active planet. However, active seismic experiments can be used to study the interiors of planets where the number of natural seismic events is not adequate for analysis.

HEAT FLOW

M. Langseth

The energy escaping from a planet in the form of heat provides one of the fundamental constraints on the nature and composition of its interior. The heat flux from the interior is comprised of several components. The two principal ones are: (1) the integrated heat production of long-lived radionuclides in the interior, which sets an upper limit on the abundance of uranium, potassium, and thorium in the planet, and (2) transient cooling of the planet since its formation, which in the two cases examined, the

Earth and Moon, appears to make a very small contribution to the present-day flux.

The very near-surface thermal regime of planets and satellites with transparent atmospheres is controlled primarily by periodic solar flux on the surface. Gradients associated with the flow of heat from the interior form only a very small component of the total flux at any given time. To reliably detect heat flow, it is necessary to do one or more of the following: (1) measure temperatures and conductivities deep in the subsurface where the effects of surface fluctuations are greatly attenuated, (2) make *in situ* measurements over a sufficiently long period to permit removal of periodic variations from the data, and (3) set up an artificial layer, on the surface, of known thermal properties (a thermal blanket) and allow it to come to equilibrium with the heat flow from below.

In situ measurements by method (1) or (2) requires that thermometers be emplaced in a vertical array in the subsurface. Sufficient penetration may be possible by hardened probes driven into the surface by free fall or rocket propulsion, provided the near-surface layers are not solid crystalline rock. Probes can also be vibrated or drilled into a powdery soil by an emplaced station. The depth of burial required for method (1) depends on the thermal properties and the period and amplitude of surface variations. On Mars, for example, a depth of about 5 m is required to avoid the annual wave, whereas on the Moon a depth greater than 0.5 m will suffice. To employ method (2), probes need not be buried as deeply. Method (3) uses a relatively thin layer with a large surface area. It has the advantage of not requiring penetration of the surface layer, but it can produce a sizable anomaly in surface temperature which will cause a long-duration transient heat flow and a steady-state distortion of the local heat-flow field. These effects must be accounted for with some precision. This method cannot be used on planets where radiative heat transfer is significant in the upper few centimeters of soil.

Measurements of heat flow made by *in situ* instruments (method 2) apply to a very small

area and consequently are subject to a number of possible near-surface disturbances, such as refraction at dipping interfaces between bodies of different conductivity, surface topography, and local variations in surface temperature. Some of these effects can be corrected for if the topography and geology are well known. Consequently, it is important to make several measurements of heat flow on a given planet to reduce effects of local distortion on the flux. In addition, global variations of heat flow are expected. Correlation of heat flow with regional variations of morphology and surface chemistry can provide constraints on the causes of these variations.

Regional variations might best be detected by remote techniques. For planets and satellites with transparent atmospheres, indications of subsurface temperatures can be obtained from measurements of microwave emissions in the range from 5 to 50 m. To determine temperatures, it is necessary to know the dielectric properties of the near-surface materials; to deduce heat flow from temperature gradients, knowledge of thermal conductivity is required. Our present knowledge of dielectric properties of possible surface materials in this range of wavelengths is scant. Further experimental data and Earth-based observations could increase confidence in the interpretation of microwave spectra. Note that radiometric measurements from the orbit of microwave emission spectra could provide global maps of the variability of heat flow over a planet's surface, even though absolute determinations may be rather inaccurate. It would therefore provide a valuable complement to a few *in situ* measurements. The applicability of microwave radiometry to the determination of heat flow varies greatly from planet to planet, depending on the electrical and thermal properties and the surface-temperature fluctuations. It would be most effective on planets with virtually no gaseous atmosphere.

For interpretation of microwave results, *in situ* measurements of near-surface thermal properties and temperature variations would be extremely valuable. Such measurements could be made by relatively simple probes on

emplaced stations that monitor the attenuation and phase shift of surface fluctuations as they propagate downward.

Lastly, subsurface temperature and thermal property determinations by means of subsurface probes are also an effective and inexpensive way to detect the existence of condensed or frozen volatiles in the soil layer.

RADAR STUDIES OF SURFACE PROPERTIES

L. Tyler, T. Howard, A. W. England, and J. Cuzzi

The study of planetary surface structure using radar techniques has been a goal of radar astronomers since the early days of radar (Evans 1969; Hagfors 1970). The scattering theory necessary to interpret reflected signals in geologic terms has reached a stage of maturity where both the limitations and capabilities of radar and the directions for future exploration are fairly well understood.

Radar studies may be divided into two modes of procedure: *monostatic*, in which the transmission and reception are done with the same antenna, and *bistatic*, in which the transmitting antenna and the receiving antenna are separated. Monostatic observations may be Earth-based (Muhleman 1966) or spacecraft-based. In the bistatic mode utilizing a spacecraft, radio signals transmitted from an orbiting vehicle are received by an Earth-based antenna after reflection from the planet's surface. Most, but not all, of the incident power is reflected specularly owing to first-order smoothness of the surface, the reflected power decreasing rapidly as the reflection angle deviates from the incidence angle. A small amount of energy is reflected into all angles, however, because of surface roughness, multiple scattering effects, or both.

This behavior has given rise to the terminology *quasi-specular component* and *diffuse component*. The quasi-specular component is the power reflected according to the Fresnel reflection coefficients, which are functions of angle and sense of polarization of the incident radiation (Stratton 1941) and the dielectric constant, ϵ , of the reflecting surface.

Determination of the dielectric constant of the reflecting surface, elevation determinations, and surface roughness estimates are the primary goals of radar studies excluding radar imaging. The dielectric constant (ϵ) of a regolith is determined by the porosity and by the intrinsic dielectric constant of the solid material (ϵ_0) of which the regolith is constructed. A good indication of the variation of ϵ_0 with rock type is given by Campbell and Ulrichs (1969), Troitskii *et al.* (1970), and Olhoeft and Strangway (1975). Typically, more basic rock types have higher dielectric constants than acidic types, and a regolith dielectric constant decreases as porosity increases. For typical regolith porosities, the variation of ϵ with rock type is not large and, consequently, in most planetary applications, remote-sensing reflectivity determinations of ϵ indicate mostly porosity variations (i.e., rock or dust).

The dielectric constants of natural silicates, oxides, and carbonates are varying functions of frequency at VHF, UHF, and microwave frequencies. Unlike complex hydrocarbons, they have no distinguishing absorption bands. Consequently, there is little hope of separating the effects caused by variations in composition from those caused by porosity by the sole use of remote-sensing techniques analogous to visible and infrared reflection spectroscopy. However, radiation in these wavelength regions is not able to penetrate the dense atmosphere of Venus. Thus the relationship between ϵ and ϵ_0 , as determined by porosity (Campbell and Ulrichs 1969; Dukhin 1971), and the dependence of ϵ_0 on rock type may make possible a potential application for Venus of a combination of radar-determined dielectric constant values and *in situ* measurements of porosity in the same surface region, perhaps by penetrators or other surface landers. The dielectric constant of water is an order of magnitude greater than that of most dry rock or soil, and water has a strong absorption band near the 3-cm wavelength. If free water is present, it will determine the microwave emissivity and strongly influence radar reflectivity.

The fact that there is diffuse scattering at regions separated from the specular point permits the construction of range-Doppler intensity maps over large regions of a planet. Such maps show clear relationships between features or units of similar scattering properties, but no completely satisfactory theoretical solution for the unambiguous determination of surface properties has been found. In a qualitative sense, however, range-Doppler mapping has produced useful images of the Moon comparable to telescopic optical photographs (Thompson *et al.* 1974). Instrumentation recently completed should produce similar, but lower, resolution images of the surface of Venus. Most of the nearside has been mapped by Earth-based radar at 100-km resolution (Rogers and Ingalls 1969; Campbell *et al.* 1970), and excellent images with resolutions in the tens of kilometers range have been produced for limited surface areas (Rumsey *et al.* 1974). These developments will soon culminate in striking images of the surface of Venus, and it is important that theoretical development leading to the geologic understanding of the signals be strongly supported.

The irregularities present in any real planetary surface also broaden or blur the sharpness of the quasi-specular peak in a way that is now well understood. Thus, in a quantitative sense, one important feature of such observations is that they permit the inference on a statistical basis of very small-scale surface structures — much smaller than those typically revealed by orbital photography — and thus permit the determination of ϵ . The geological worth of such data when combined with images of the surface has been demonstrated for the moon where multi-frequency observations have been used to categorize geologic units and may be indicators of relative ages of the units (Moore *et al.* 1975).

In all spacecraft-implemented bistatic observations, power and antenna gain limitations permit sufficiently sensitive measurements of only the quasi-specular component where most of the

reflected power lies. This component of the scattering originates from a small region around the point of reflection on the mean surface. The location of this point moves as the position of the spacecraft changes along its orbit, giving rise to a scanning action that permits study of large regions on the Earth side of the planet. Successful systematic observations have been carried out using bistatic spacecraft techniques for the Moon (Tyler 1968; Tyler and Simpson 1970; Tyler *et al.* 1971; Tyler and Howard 1973) and for Mars (Fjeldbo *et al.* 1972). In the case of monostatic Earth-based observations of the specular component, the scattering arises from a small region surrounding the sub-Earth point. For these observations, the scanning action originates from the relative motions of the Earth and the planet and is generally limited to equatorial regions. This type of observation has been useful for Mars where rapid rotation and large distance produce a sensitivity limited situation (Downs *et al.* 1973; Pettengill *et al.* 1969, 1973; Evans 1969). By observing the specular component of the sub-Earth region with a resolution of several kilometers, variations in reflectivity and surface roughness may be mapped. Simultaneous range measurements, in which the return time of a pulsed signal is measured, allow altitude variations to be mapped for the same region (Goldstein *et al.* 1970). Using a correlation of observations of this type, Pettengill *et al.* (1973) suggest that the higher regions of Mars were more compacted and the lower regions were covered with relatively thick (meters) regolith layers. The earliest quantitative radar determinations of slope and altitude variations showed that Mars is topographically interesting and has significant relief (~ 15 km). More recent comparisons of Mariner 9 imagery with Earth-based radar observations (Simpson *et al.* 1974) clearly demonstrate the sensitivity of radar-reflectance to wavelength-scale surface structure.

REFERENCES CITED

Campbell, D. B., Jurgens, R. F., Dyce, R. B., Harris, F. S., and Pettengill, G. H., 1970. Radar interferometric observations of Venus at 70-centimeter wavelength. *Science* 170:1090-1092.

Campbell, M. J., and Ulrichs, J., 1969. Electrical properties of rocks and their significance for lunar radar observations. *J. Geophys. Res.* 74:5867-5881.

Dainty, A. M., Toksoz, M. N., Solomon, S. C., Anderson, K. R., and Goins, N. R., 1974. Constraints on lunar surface. Proc. Fifth Lunar Science Conf., *Geochim. Cosmochim. Acta*, Suppl. 5, pp. 3091-3114.

Downs, G. S., Goldstein, R. M., Green, R. R., Morris, G. A., and Reichley, P. E., 1973. Martian topography and surface properties as seen by radar: the 1971 opposition. *Icarus* 18:8-21.

Dukhin, S. S., 1971. Dielectric properties of disperse systems in *Surface and Colloid Science*, E. Matijevic, ed., New York:Wiley-Interscience, pp. 83-86.

Dyal, P., Parkin, C. W., and Daily, W. D., 1974. Magnetism and the interior of the Moon. *Rev. Geophys. Space Phys.* 12:568-591.

Evans, J. V., 1969. Radar studies of planetary surfaces, ed. by Goldberg, L., Layzer, D., and Phillips, J. G. *Ann. Rev. Astronomy Astrophys.* 7:201-248.

Fjeldbo, G., Kliore, A., and Seidel, B., 1972. Bistatic radar measurements of the surface of Mars with Mariner 1969. *Icarus* 16:502-508.

Fuller, M., 1974. Lunar magnetism. *Rev. Geophys. Space Phys.* 12:23-70.

Goldstein, R. M., Melbourne, W. G., Morris, G. A., Downs, G. S., and O'Handley, D. A., 1970. Preliminary radar results of Mars. *Radio Sci.* 5:475-478.

Hagfors, T., 1970. Remote probing of the Moon by infrared and microwave emissions and by radar. *Radio Sci.* 5:189-227.

Kaula, W. H., 1968. *An Introduction to Planetary Physics: the Terrestrial Planets*. New York: John Wiley and Sons.

Lin, R. P., McGuire, R. E., Howe, H. C., and Anderson, K. A., 1976. Mapping of the lunar surface remanent magnetic fields by electron scattering. Proc. Sixth Lunar Science Conf., *Geochim. Cosmochim. Acta* Suppl. 6, 3:2971-2973.

Moore, H. J., Tyler, G. L., Boyce, J. M., Shorthill, R. W., Thompson, T. W., Walker, A. S., Wilhelms, D. E., Wu, S. S. C., and Zisk, S. H., 1975. Correlation of photogeology and remote sensing data along the Apollo 14 bistatic-radar ground track: a working compendium. Interagency Rept. Astrogeology 75, U.S. Geol. Survey open file report.

Muhleman, D. O., 1966. Planetary characteristics from radar observations. *Space Sci. Rev.* 6:341-364.

Murray, B. C., 1975. Mercury. *Scientific Am.* 233(3):58-68.

Olhoeft, G. R., and Strangway, D. W., 1975. Dielectric properties of the first 100 meters of the Moon. *Earth and Planet. Sci. Ltrs.* 24:394-404.

Pettengill, G. H., Counselman, C. C., Rainville, L. P., and Shapiro, I. I., 1969. Radar measurements of martian topography. *Astronom. J.* 74:461-482.

Pettengill, G. H., Shapiro, I. I., and Rogers, A. E. E., 1973. Topography and radar scattering properties of Mars. *Icarus* 18:22-28.

Rogers, A. E. E., and Ingalls, R. P., 1969. Venus: mapping the surface reflectivity by radar interferometry. *Science* 165:797-799.

Rumsey, H. C., Morris, G. A., Green, R. R., and Goldstein, R. M., 1974. A radar brightness and altitude image of a portion of Venus. *Icarus* 23:1-7.

Stratton, J. A., 1941. *Electromagnetic Theory*. New York:McGraw-Hill Book Co.

Thompson, T. W., Masursky, H., Shorthill, R. W., Tyler, G. L., and Zisk, S. H., 1974. A comparison of infrared, radar, and geologic mapping of lunar craters. *The Moon* 10:87-117.

Troitskii, V. S., Bondar, L. N., Zelinskaya, M. R., Strezhneva, K. M., 1970. Comparison of the chemical composition of lunar surface material determined by radioastronomical observations with the results of chemical analysis obtained by Surveyor. *Radio Sci.* 5:247-251.

Tyler, G. L., 1968. Oblique-scattering radar reflectivity of the lunar surface: Preliminary results from Explorer 35. *J. Geophys. Res.* 73(24):7609-7620.

Tyler, G. L., and Simpson, R. A., 1970. Bistatic radar measurements of topographic variations in lunar surface slopes with Explorer 35. *Radio Sci.* 5:263-271.

Tyler, G. L., and Howard, H. T., 1973. Dual frequency bistatic radar investigations of the Moon with Apollos 14 and 15. *J. Geophys. Res.* 78:4852-4874.

Tyler, G. L., Simpson, R. A., and Moore, H. J., 1971. Lunar slope distributions: A comparison of bistatic radar and photographic results. *J. Geophys. Res.* 76:2790-2795.

CHAPTER V. GEODESY AND CARTOGRAPHY

R. Batson and D. Arthur

Geodesy involves the determination of the size and shape of a planet and the location of points on its surface. It also involves the study of gravitational fields and temporal variations such as planetary rotation. Cartography involves map-making operations, from the acquisition of appropriate study data through the final printing of the map. Geodesy and cartography provide the geometric framework on which most investigations of planets are ultimately based. Specifically, the products of these disciplines provide information on the following:

1. The dimensions of the planet
2. A mathematical figure of reference for the planet
3. The orientation of the body in the celestial coordinate system
4. The rotational constants
5. A defined system of coordinates
6. The location of surface points in the defined coordinate system
7. The gravity potential expressed in spherical harmonics
8. Topographic and thematic maps:
 - a. At small scales for synoptic coverage
 - b. At intermediate scales for regional coverage
 - c. At large scales for spacecraft landing areas and other sites of particular scientific value
9. Surface albedo in various wavelengths

In this chapter, the relevance of geodesy and cartography to planetology is discussed and the

requirements of data acquisition and mission design are considered. Many of the techniques used in planetary geodesy and cartography were derived from mapping the Moon, as discussed by Kopal and Carder (1974).

GEODESY

The literal meaning of geodesy is to “divide the Earth.” In application, it has come to mean the determination of the size and shape of the physical Earth and the location of a network of control points on the Earth’s surface in fixed coordinate systems. Because of the instrumental techniques used for measurements on the Earth’s surface, geodesists have had to concern themselves with the positions of celestial bodies and with the gravity field. Their activities and interests overlap those of astronomers and geophysicists and, as a consequence, geodesy has been divided into three nearly distinct subjects: geometric geodesy, geodetic astronomy, and gravimetric (physical) geodesy. Detailed discussions of these aspects of geodesy can be found in Bomford (1962).

Generally, the size, shape, and rotational constants contain geophysical information about moments of inertia, internal composition, and rigidity. The harmonics of the gravity field contain information about anomalous mass distribution within the body (see chapter entitled “Geophysics”). Geodesy provides coordinate systems

and point locations that are necessary to correlate all other data. In addition, geodesy provides the control for site locations and for navigation and guidance of landers and surface-mobile vehicles. Geodetic studies are therefore essential for the study of any planetary body with a solid surface.

For most of its history, geodesy has had to depend on observations made from the surface of the Earth. The restricted range of these observations made possible the preparation of highly detailed and precise maps of localized areas, but expansion of dimensions and reference systems to continental and global areas was accomplished only with great difficulty, and some unresolved ambiguities still exist. The capability of observing from and to artificial satellites, available only in the past 15 years, has completely revolutionized geodesy. When applied to the planets, these techniques make feasible the scientifically preferable approach of proceeding from the general to the specific in an orderly fashion.

CARTOGRAPHY

Cartography, or map-making, involves three main phases: (1) data acquisition, (2) computation of controls, and (3) map compilation. Traditionally, these phases are performed in serial order, but the pressures of spaceflight mission operations and of short-lived project funding requires that these three phases progress almost simultaneously.

Data Acquisition

Data acquisition is the single most important phase because it generally cannot be repeated, and lost or non-useful planetary data are nearly impossible to recover later. For this reason, on missions in which interaction with the spacecraft data-gathering systems is possible, much preliminary cartographic work is performed in "near real-time" to ensure that planned sequences are actually gathering adequate information to map the planet. The requirements of data acquisition systems are discussed below.

Computation of Controls

Computation of controls for making maps of planetary surfaces is derived from geodetic analyses that include definition of coordinate systems (deVancouleurs *et al.* 1973), derivation of the gravitational figure of the planet as data for map projections and elevation reference (Jordan and Lorell 1973), and computation of latitudes and longitudes of a set of identifiable map images to which all map features are referenced (Davies and Arthur 1973; Davies and Batson 1975). The latter is performed by analytical photogrammetric methods.

In fact, planetary mapping relies heavily on photogrammetry, the science of measuring objects from photographic images. The measurements generally utilize stereopsis, wherein an object is viewed simultaneously from two points of a perspective so that two two-dimensional images are fused into a single three-dimensional image. Photogrammetric instruments normally consist of (1) a binocular viewing apparatus for examining stereoscopic pairs of pictures, (2) a cursor, or "floating mark," appears to the operator to float in the three-dimensional image, and (3) an analog (mechanical) or digital computer linked to a plotting table on which a stylus traces the x and y coordinates of the "floating mark" in the stereoscopic image viewed by the operator. The "z", or vertical, coordinate of the floating mark is visible to the operator as a number on a mechanical or electronic counter. When used with high quality photographs taken specifically for stereophotogrammetry, extraordinary accuracies are possible. Horizontal locations of features can be determined within the limits of resolution of the image. Measurement of vertical positions within 1/10,000 of the distance between the camera and the surface being mapped is common.

Analytical photogrammetry is a mathematical technique for computing a relatively small set of locations of objects as a function of the coordinates of their images on two or more pictures. Stereopsis is used primarily to identify conjugate images, but not as an essential part of the measuring process. The computation is not limited

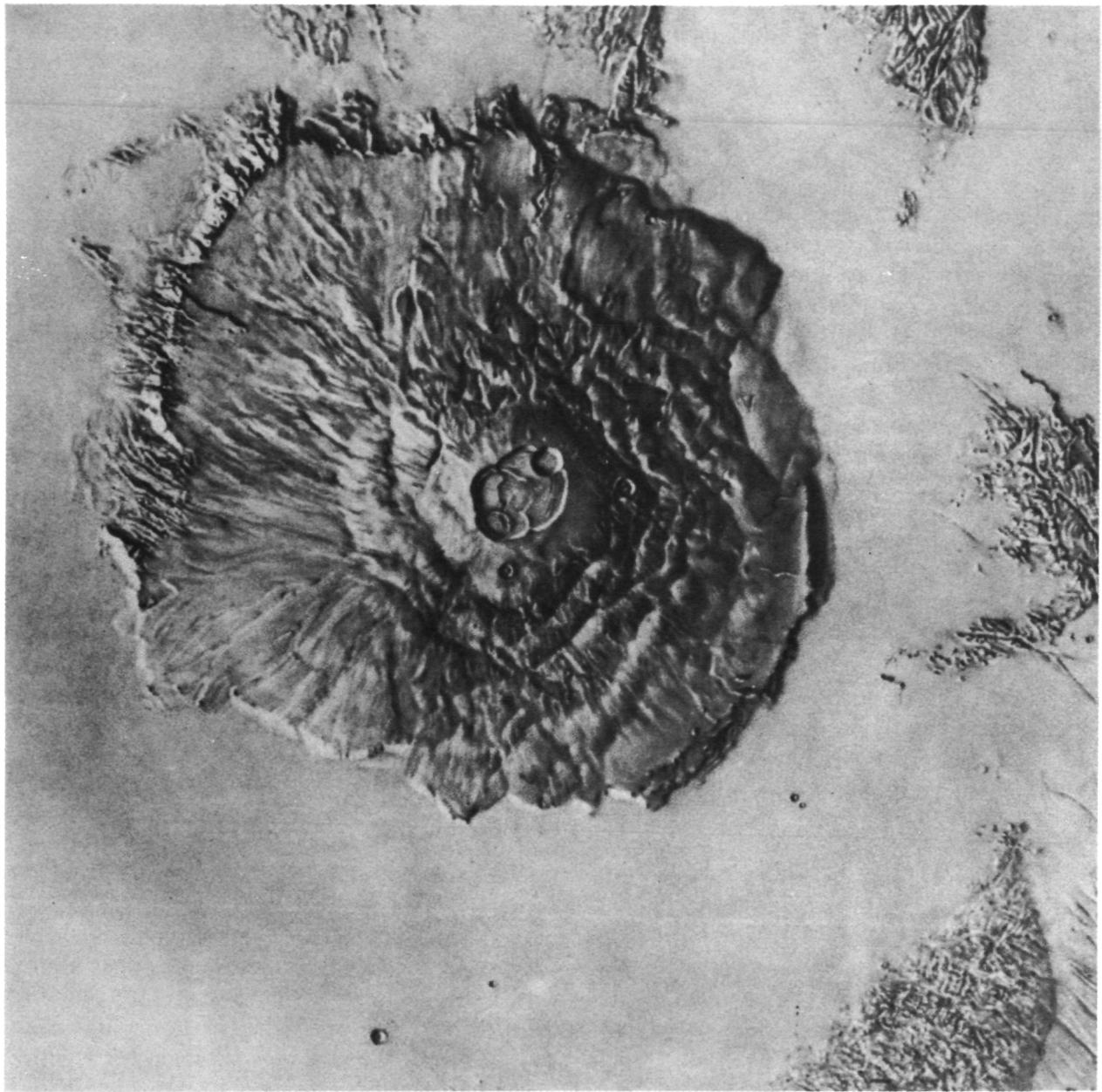


Figure 5.1. Airbrush rendition of Olympus Mons, the ~600 km-diameter shield volcano on Mars. Maps of planetary surfaces typically are portrayed either as photomosaics or as airbrush renditions. Each product serves unique functions. Airbrush versions show the terrain under uniform lighting, and contrast can be enhanced or subdued as needed. For example, as a base for geological maps, the terrain should be subdued; on the other hand, topical studies of specific areas would require a more prominent version, as shown here. Images from which the airbrush versions are prepared are often obtained with lighting from the opposite direction than desired. Consequently, regions that will be "lighted" in the airbrush version may actually be in shadow, but must be portrayed. Computer-processed photomosaics show the maximum detail and are therefore more useful for compilation of data and interpretation of results. Irregular lighting (often from the "wrong" position) and nonuniform resolution pose problems for use as final products. (Courtesy of U.S. Geological Survey.)

to human physiological constraints, and any number of overlapping pictures may be used for optimum determination in the computations.

Map Compilation

Map compilation includes the geometric transformation of pictures to appropriate map projections, the assembly of photomosaics, and final rendition as finished map products (Fig. 5.1) by illustrators and draftsmen (Batson 1973a,b). Where appropriate data exist, it also includes the compilation of contour lines by stereophotogrammetry (Wu *et al.* 1973) or by interpolation between spot elevation control points derived by other methods (Christensen 1975; Wu 1974).

A limited set of map scales differing by factors of 4 or 5 can be devised to portray all useful cartographic detail on a planet. Limiting the number of scales reduces the cost of map plotting and general formatting and, more importantly, reduces confusion involved in the study of geologic features. For example, maps of a volcanic field at slightly different scales may induce misleading comparisons, whereas if the scales are either exactly the same or distinctly different, then the true size, structures, and textures of features will be immediately apparent.

For the Moon, a map series was recommended by the Geology Panel at the 1967 Conference on Lunar Science and Exploration at Santa Cruz, as follows:

Orthographic, Mercator, and polar stereographic projections of the whole Moon at a scale of 1:5,000,000.

Complete topographic coverage of the Moon at a scale of 1:1,000,000.

Coverage of approximately 20 areas of scientific interest for landing sites and traverses at a scale of 1:250,000.

Coverage of central parts of 20 areas of special interest at a scale of 1:50,000.

Coverage of landing locations in the central part of the areas of science interest at a scale of 1:5,000.

For Mars and Mercury, the following series were selected by experiment team members of Mariners 9 and 10:

Mercator projection from 65°N to 65°S and polar stereographic projections north and south of 55° at a scale of 1:25,000,000 for the entire planet.

Mercator, Lambert conformal, and polar stereographic projections of the entire planet at 1:5,000,000.

Selected areas of special scientific interest (including spacecraft landing sites) at 1:1,000,000 on transverse Mercator projections.

Selected areas of special scientific interest at 1:250,000 on transverse Mercator projections.

These map series appear to be useful for maps of any planet likely to be mapped in the foreseeable future and satisfy most of the requirements for map bases in geologic mapping (see chapter entitled "Stratigraphy and Structure").

Under ideal conditions, map accuracy requirements would be as shown in Table 5-1.

These requirements assume an abundance of data, permitting rejection of excessive detail. Such a wealth of material has not yet become available to planetary cartographers, so they have taken the alternate course of selecting the smallest map scale that will accommodate the highest density of detail available (other than that in specific small areas such as Mariner 9 B-camera pictures) and using that scale for an entire series. In the mapping of Mars, for example, scales were selected that would permit uncluttered cartographic portrayal of all available pertinent material larger than 0.5 mm in diameter at map scale, whether horizontal and vertical standard errors were commensurate with the map resolutions or not. The resolution of map data by this criterion was high enough to justify systematic 1:5,000,000 mapping in the southern hemisphere of Mars, but not in the north. The 1:5,000,000 scale was nevertheless retained in the north for consistency.

DATA-ACQUISITION SYSTEMS

The basic data for planetary geodesy and cartography are provided mostly by spacecraft

TABLE 5.1

| Map scale | Horizontal standard error, m | Optical resolution, m | Contour interval, m | Vertical standard error, m |
|-----------|------------------------------|-----------------------|---------------------|----------------------------|
| 5,000,000 | 1,520.0 | 250.00 | 1,000 | 300 |
| 1,000,000 | 300.0 | 50.00 | 500 | 150 |
| 250,000 | 76.0 | 12.50 | 100 | 30 |
| 50,000 | 15.0 | 2.50 | 25 | 8 |
| 5,000 | 1.5 | .25 | 3 | 1 |

imaging systems and are reduced by methods of photogrammetry. Given unlimited resources, a photogrammetrist would request that images to be used in planetary mapping be taken with wide-angle lenses ($50\text{-}120^\circ$ frame diagonals), that they be recorded on dimensionally stable film at least 100 mm wide and with resolution of at least 50 lines/mm. He would further request that the image of a star field, or other data for determining precise camera altitude, be recorded at the instant the cartographic picture is taken and that all film be returned to Earth. Although such a system has been used during the late Apollo missions for lunar mapping, its weight and sophistication (including the requirement that film taken in orbit around a planet be returned to Earth) make the system impractical for planetary mapping, at least for the present. Since each element of the above specifications must be compromised to some degree in planning planetary missions, it is important to consider the reasons for these ideal specifications when planning alternative approaches. Failure to do so can have serious consequences. For example, the Lunar Orbiter program was designed to provide surface photographic coverage at medium and high resolution. That function was performed very well, but image geometry in Lunar Orbiter pictures has such large random variations that when landmark positions, elevations, terrain slopes, and control geometry were required, the system was grossly inadequate.

If the preferred imaging system specified above is used, camera position and altitude can be determined photogrammetrically more accurately than by any other means. For wide-

angle photography, the necessary conditions for reconstitution are that five points must be identified on two photographs and that no three of these should lie in one plane containing the two camera positions (Fig. 5.2a) – this is the well-known Fourcade correspondence theorem. Base:height ratios of 0.2-1.5 are usable, 1.0 being about the optimum. The Fourcade correspondence theorem cannot be applied to narrow-angle photography, but pure photogrammetric reconstitution is possible if a minimum of three pictures with considerable convergence ($<30^\circ$) is acquired, with the same four points measured and identified on all three pictures (Fig. 5.2b). Reconstitution is possible from two strongly convergent photographs provided accurate orientation and positional information on the camera is available from spacecraft-tracking data, so that the correspondence settings can be introduced (rather than derived), and provided the camera's internal geometry is appropriately calibrated. Other geometric conditions are illustrated in Figures 5.2c and 5.2d. Despite their disadvantages for geodetic purposes, narrow-angle cameras have been used on previous lunar and planetary missions and will continue to be used on planetary missions presently planned because they allow acceptable ground resolutions to be achieved from normal orbital altitudes.

The critical point in planetary mapping is that the image geometry of the data-acquisition system be accurately calibrated. That is, it must be possible to define mathematically the geometry of images gathered by the system. Systems in which the geometric parameters of acquired

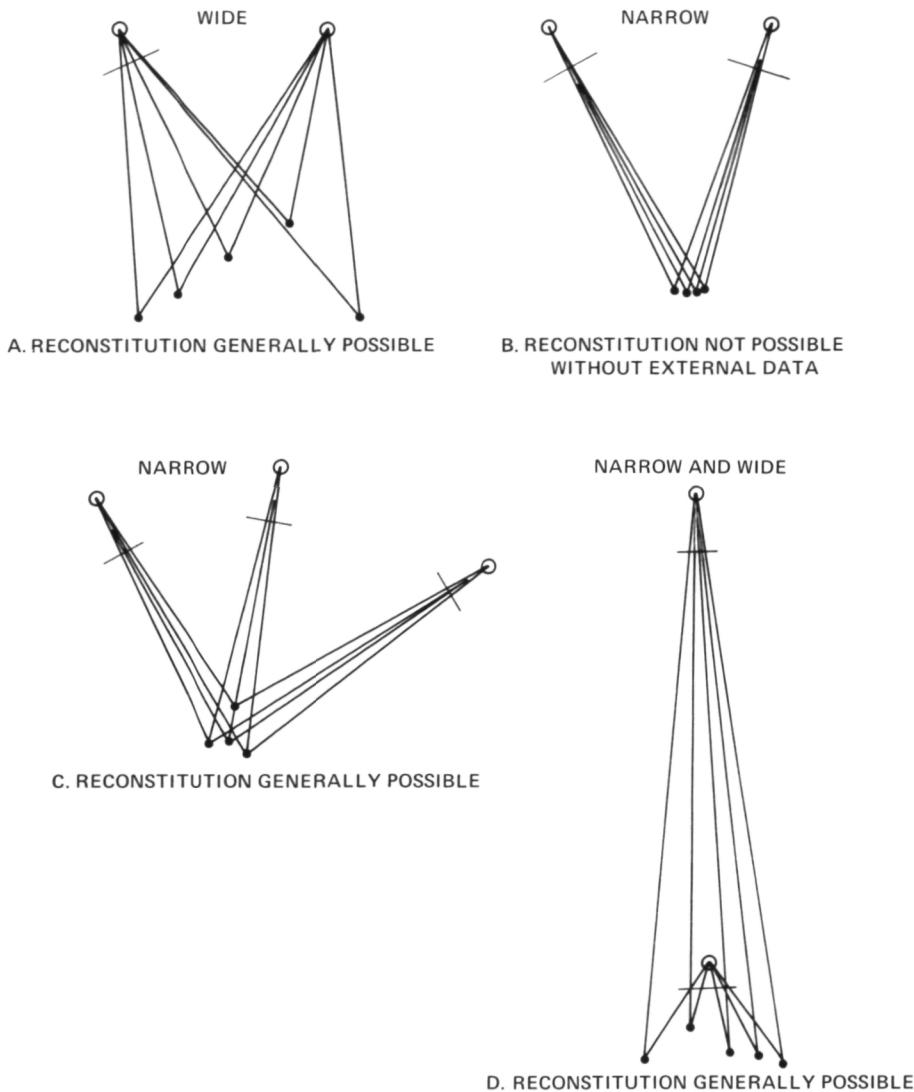


Figure 5.2. Various spacecraft imaging arrays related to photogrammetric reconstitution.

images are not precisely repeatable should not be used in planetary exploration.

For planets with atmospheres that are transparent to visible wavelengths, conventional images will be the basis for most of the cartography. Generally, relative elevations and positions can be obtained satisfactorily from visual images, but absolute values are difficult to obtain. For Mars, terrestrial radar and spacecraft occultations have proved invaluable for calibrating the ultraviolet spectrometer elevation data. They could perform a similar control function for photographic data once extensive stereo is acquired, as was done for a study of Olympus

Mons on Mars by Davies (1974). For planets with atmospheres that are opaque in the visible wavelengths, radar will be the only means to obtain cartographic data. Fortunately, radar and visual data can be portrayed in essentially the same manner. Discrete radar elevation measurements and profiles, such as would be acquired from terrestrial observations, can be integrated and portrayed as shaded relief with any chosen Sun angle. Computer programs already exist for doing this routinely. Similarly, images from side-looking radar can be readily transformed into any appropriate map projection and used as a cartographic base.

In addition to radar, other non-imaging sensors can provide important geodetic and cartographic data. For example, atmospheric pressures measured by infrared and ultraviolet spectroscopy provided valuable surface elevations for mapping Mars with Mariner 9 (Hord *et al.* 1974; Conrath *et al.* 1973). Spacecraft tracking provides data required for determination of an equilibrium surface (or "topographic datum") to which surface elevations must be related if they are to be meaningful to geologists and geophysicists. Generally, however, all non-image data must be related to a much more comprehensive set of image data before detailed cartography can begin. Even for Venus, where preliminary mapping is being done with radar soundings resulting in terrain elevations, a sufficient number of points must be measured to allow creation of an image of the surface if meaningful maps are to be made.

MISSION DESIGN

Geodesy

Both orbital and flyby photography are useful for geodetic purposes. Approach photography that encompasses all or most of the disk can be very important. In addition, the photographic period should be extended to provide continuity in scale between full-disk and detailed photography of the surface. The main aims of approach photography are:

1. Measurement of the profile of the disk to determine ellipticity
2. Determination of the relative three-dimensional positions of surface markings
3. Fitting of the surface markings to an oblate spheroid and fixing the axis of rotation with respect to surface markings
4. Determination of the position of the axis of rotation with respect to the celestial sphere

The aims are pertinent to all sizable bodies in the solar system except those that have no solid surface. Two general situations are encountered: rapidly rotating planets (e.g., Mars) and relatively slow-rotating planets (e.g., Mercury). For

rapidly rotating planets, it is possible to make good use of long-focus imaging systems that are generally desirable in other fields of study. Long-focus systems depend on the rotation of the planet to provide a baseline. In this case, the inclination of the vehicle trajectory to the plane of the equator is important. Clearly, if the spacecraft is on a polar axis during approach, then as viewed from the surface of the planet, it is not displaced by rotation. Instead, the position of the spacecraft remains fixed and a three-dimensional triangulation is not possible. To utilize the planet's rotation, the trajectory must be in a plane near the equatorial plane of the planet.

To establish a baseline for triangulation on a slowly rotating planet, the rotation is of no assistance and image displacement must be caused entirely by spacecraft motion. Since this motion is almost directly toward the planet, the position in the sky of the spacecraft, as seen from the planet, will change slowly and poor geometry will result until near encounter. At encounter, the direction of spacecraft motion is at a significant angle to the line joining the spacecraft and the center of the planet, and so the spacecraft moves rapidly across the sky and good geometry is possible.

A similar geometric problem is encountered in establishing the position of a rotation axis with respect to the celestial sphere from approach photography. Since the approach trajectory is nearly a straight line, the position of the spacecraft on the line alone is insufficient to define the geometry because the planet and the cameras can be rotated as one around the trajectory without violating the tracking data or the photographic measures. To rigidly define the geometry, the orientation of the camera must be known. The best way to obtain this orientation is by simultaneous photography of the star field with a secondary camera with a wide-angle lens.

Map Control and Cartography

Uniform distribution of map data over the entire planetary surface is the ultimate goal of planetary mapping. This goal remains unrealized,

even for the Earth. The traditional approach to planetary exploration is to send a spacecraft with an imaging system on a trajectory as a flyby that takes it near the planet or planets under investigation. The flyby missions are then followed by more expensive and sophisticated orbiters and finally with landers. Maps can be made with data from all of these missions, but extreme variation in data resolution in flybys or orbiters with highly elliptical orbits can result in misleading maps because, at a given scale, they depict data with variable resolution. For example, the footprint of a Mariner 9 picture element is about 0.5 km square at lat 20°S on Mars, but is about 1.5 km square at lat 60°N. Yet the 1:5,000,000-scale Mars map series portrays *all* areas as if the density of map data were uniform over the entire planet. Furthermore, widely different mission sequences often must be used to gather various kinds of data. For example, albedo data should be gathered from high-altitude with high-Sun illumination. Computation of a net of horizontal control points is best done with high-altitude, low-Sun pictures. Planimetric maps of surface detail are best made with monoscopic low-altitude pictures taken under low-Sun conditions, whereas surface detail to be mapped stereoscopically should be photographed at low-altitude under intermediate (20-40°) illumination. Obviously, all of these different requirements cannot be met during a single mission and various compromises must be derived.

If it were not for conflicts with other experiments, cartographic data for albedo and planimetric maps could be gathered fairly easily with existing Mariner-type spacecraft. Ideally, two spacecraft would be used. The first would be placed in a high-altitude, high-inclination orbit synchronized with the rotation of the planet to permit high-altitude, high-Sun angle and high-altitude, low-Sun coverage of the entire planet. The second spacecraft would be placed in a polar orbit synchronized with the planetary rotation in such a way that overlapping strips of high-resolution pictures could be taken near the terminator during each revolution until high-resolution coverage of the entire planet were obtained. This approach was, in fact, the one

attempted with the Mariner Mars 1971 mission, but the loss of a spacecraft and the presence of a planet-wide dust storm required that mission sequence compromises be made that were fully acceptable to no one.

DATA REDUCTION

The camera focal lengths and picture formats useful for planetary exploration from spacecraft, whether flyby, orbiter, or hard- or soft-lander, are different from those used in conventional terrestrial mapping and they are more or less incompatible with the photogrammetric procedures and equipment generally available. In addition, information about camera position and attitude — available through orbital tracking and attitude sensors or star pictures — provides geometric constraints on triangulation not available to usual aircraft situations. This means that computer programs designed for conventional aerial control extension are inadequate for planetary missions. Consequently, consideration must be given, from the initiation of a planetary exploration program through complete system design, to not only the spacecraft, the sensors, and the mission, but also to the hardware and software for extracting the maximum geodetic and cartographic information from the records obtained. Unfortunately, this has not always been the case through the lunar program and the planned planetary flights.

SUMMARY

1. Geodesy and cartography provide a reference system for navigation and guidance and for correlation and compilation of all data about the surface of planets and satellites.

2. The basic requirement for geodesy is geometrically true, low-to-moderate-resolution, wide-angle photography of the whole planet. For cartography, the basic requirement is high-resolution, wide-angle photography of areas of interest, the resolution depending on the intended map scale.

3. Resolution demands and weight constraints have forced and are likely to continue to force

the use of narrow-angle camera systems for which reconstitution requires accurate knowledge of external data, such as camera orientation and position. The narrow-angle cameras demand more stringent calibration procedures, a more restrictive imaging strategy, and more precise knowledge of camera orientation and position than wide-angle cameras; these factors must be considered in total system design at the initiation of any imaging experiment.

4. The most necessary attribute of the imaging system, without which no metric analysis is possible, is geometric fidelity. The widely used vidicon systems are notoriously prone to

distortion. New imaging systems, under development for planetary missions, should be designed to ensure geometric reliability.

5. Useful photogrammetric determinations of elevations with narrow-angle cameras from orbital altitudes is possible only with convergent stereophotography, and then only with precisions of the order of several hundred meters for most of the camera systems now being considered for planetary exploration.

6. Nonvisual data can be used for geodetic and cartographic purposes, and with radar, the additional ranging capability can be used for elevation measurements.

REFERENCES CITED

Batson, R. M., 1973a. Cartographic products from the Mariner 9 mission. *J. Geophys. Res.* 78:4424-4435.

Batson, R. M., 1973b. Television cartography. U.S. Geol. Surv. Interagency Rept. Astrogeology 58.

Bomford, G., 1962. Geodesy, 2nd edition. Oxford:Clarendon Press, 561 p.

Christensen, E. J., 1975. Martian topography derived from occultation, radar, spectral, and optical measurements. *J. Geophys. Res.* 80:2909-2913.

Conrath, B., Curran, R., Hanel, R., Kunde, V., Maguire, W., Pearl, J., Pirraglia, J., Welker, J., and Burke, T., 1973. Atmospheric and surface properties of Mars obtained by infrared spectroscopy on Mariner 9. *J. Geophys. Res.* 78:4267-4278.

Davies, M., 1974. Photogrammetric measurements of Olympus Mons on Mars. *Icarus* 21:230-236.

Davis, M. E., and Arthur, D. W. G., 1973. Martian surface coordinates. *J. Geophys. Res.* 78:4355-4394.

Davies, M. E., and Batson, R. M., 1975. Surface coordinates and cartography of Mercury. *J. Geophys. Res.* 80:2417-2430.

deVaucouleurs, Gerard, Davies, M. E., and Sturms, F. M., Jr., 1973. Mariner 9 aero-graphic coordinate system. *J. Geophys. Res.* 78:4395-4404.

Hord, C. W., Simmons, K. E., and McLaughlin, L. K., 1974. Mariner 9 ultraviolet spectrometer experiment: Pressure-altitude measurements on Mars. *Icarus* 21:292-302.

Jordan, J. F., and Lorell, J., 1975. Mariner 9, an instrument of dynamical science. *Icarus* 25:146-165.

Kopal, Z., and Carder, R. W., 1974. Mapping of the Moon. Boston:Reidel, 237 p.

Wu, S. C., 1974. Topographic mapping of Mars. U. S. Geol. Surv. Interagency Rept. Astrogeology 63.

Wu, S. C., Schafer, F. J., Nakata, G. M., Jordan, R., and Blasius, K. R., 1973. Photogrammetric evaluation of Mariner 9 photography. *J. Geophys. Res.* 78:4405-4410.

CHAPTER VI. EARTH-BASED STUDIES

S. Dwornik

The goals of planetary geology are to understand the origin, evolution, distribution, composition, and interrelationships of condensed matter in the form of comets, asteroids, and planets and their satellites. These goals are directly related to the basic objectives of space exploration discussed in the introductory chapter of this report.

To achieve these goals, it is necessary to understand the basic geologic processes that have been involved in the evolution of planetary bodies. From previous lunar and planetary missions, it is now clear that we are dealing with essentially the same geologic processes on all planets, but the *intensity* and *time* factors of individual processes vary with the environment of the planet; at the same time, environment itself often varies as a function of surface processes. To understand the general processes, it is necessary to perform a variety of tasks involving laboratory experiments and simulations, theoretical modeling studies, telescopic observations, and comparative geological field work. These types of studies are required to expand our fundamental knowledge of the planets and to help interpret space-mission data that might appear to be ambiguous or seemingly in conflict with past observations or data. Theoretical understanding of geologic processes of all planets and satellites is the ultimate goal of comparative geologic studies, simply because it is not possible to sample and study all bodies in the solar system *in situ*. Theories must make verifiable

predictions, be able to test the aptness of terrestrial analogs, and ultimately should provide direction for future planetary research.

For example, let us examine the general process of wind erosion which can be present wherever an atmosphere and solid surfaces exist. On Earth, wind is a rather effective erosive process in dry climates. Without present knowledge of the near-surface atmosphere conditions on Venus, we cannot say with certainty if wind erosion is occurring there, but it appears to be within the realm of possibility. Even before Mariner-Mars, telescopic observations of Mars led many investigators to conclude that wind processes were active on the martian surface. Mariner 9 confirmed these observations by revealing numerous features that seemed to be directly linked to the great dust storm of 1971. Most of these features were associated with craters and many of them were observed to change in shape, size, and position as the dust storm subsided. In fact, wind erosion appears to be one of the dominant surface processes on Mars acting in the recent geologic history.

Thus, we are concerned with the same general geologic process — wind erosion — operating under a variety of conditions on three planets. The atmospheric pressures range over about five orders of magnitude (Venus surface pressure ~ 100 bar, Earth ~ 1 bar, Mars ~ 0.005 bar). To understand the nature of wind erosion on any one planet, it is first necessary to understand wind erosion in general by knowing the

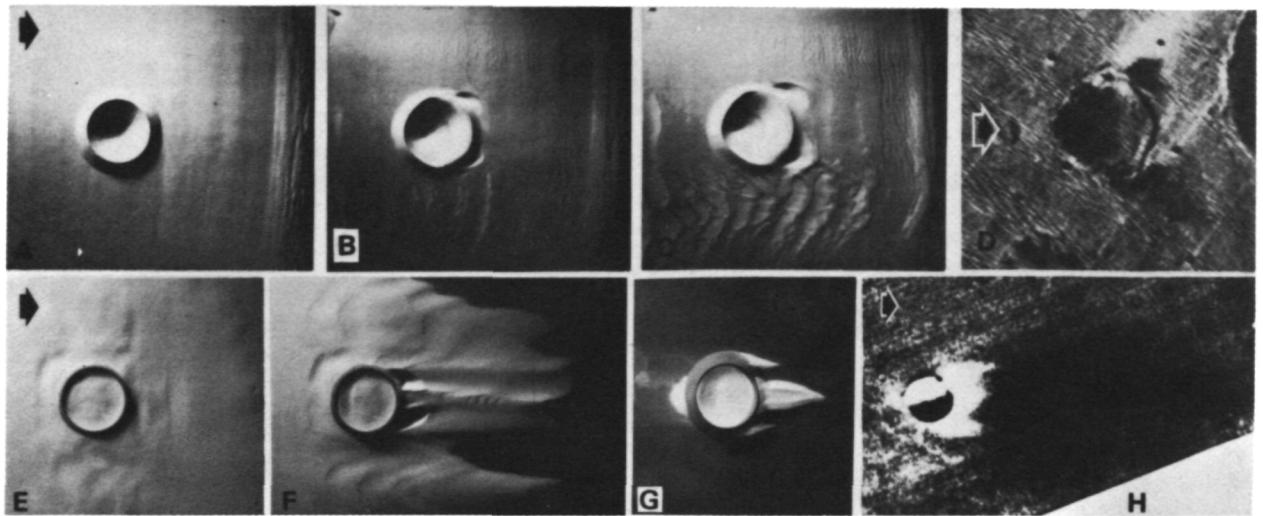


Figure 6.1. Wind-tunnel simulations of flow over a raised-rim crater to determine zones of aeolian erosion and deposition, and comparisons with aeolian phenomena on Mars. (A to C) Sequential photographs (light from upper left, wind from left to right as indicated by the arrow) of a 17.8-cm crater modeled in loose sand and placed in the wind tunnel with a wind velocity of about 420 cm/sec. The crater became ovoid in outline, pointing upwind and developed two erosional depressions corresponding to the trailing components of a horseshoe vortex. (D) Small martian crater in Mare Tyrrhenum showing similar outline and dark zones off the leeward edge of the crater rim that are interpreted to be the result of erosion. (E to G) Sequential photographs of a 17.8-cm crater modeled in solid wood (nondeformable), partly buried by loose sand (E) and subjected to a 850-cm/sec wind until relatively stable conditions ensued (G) in which the model surface was swept free of loose sand except in zones of relative deposition, shown by the white trilobate pattern and the white patch on the windward rim. (H) A 2-km martian crater in the region northwest of Memnonia showing a similar trilobate pattern in the immediate lee of the crater and a white zone on the windward rim that are interpreted to be aeolian deposits, and a large dark zone in the crater wake area interpreted to have resulted from erosion (Mariner 9 shading-corrected image) (from Greeley et al. 1974).

parameters involved and the effects of varying the values of those parameters.

Laboratory wind-tunnel simulations provide the opportunity to single out the parameters involved in wind erosion and to vary their magnitude to determine their relative importance. Thus the fundamental process of wind erosion can be better understood. Next, the environmental planetary conditions can be duplicated (to some degree) to physically model the particular planet under study. For example, the Martian Surface Wind Tunnel at NASA-Ames Research Center developed by Greeley permits particle threshold experiments to be conducted at true martian pressures.

In addition to parametric analyses, modeling studies in the wind tunnel enable the determination of the flow field over various topographic

models, such as craters and dunes, to determine zones of wind erosion and deposition. A first-order determination has been completed for raised-rim craters which appears to explain certain dark and light streaks associated with martian craters (Fig. 6.1). Although certain modeling laws would indicate that these results are valid, a natural reluctance on the part of some investigators is understandable in comparing 30-cm wind-tunnel models with full-size (1 km and larger) craters. In this respect, at least to a qualitative degree, field studies of full-size craters are important. Large natural and man-made craters having associated aeolian features are currently being sought as a test of the modeling results obtained in the wind tunnel. Several examples have been found. Wolf Creek Crater (Australia), Tememichat Crater (Africa), and

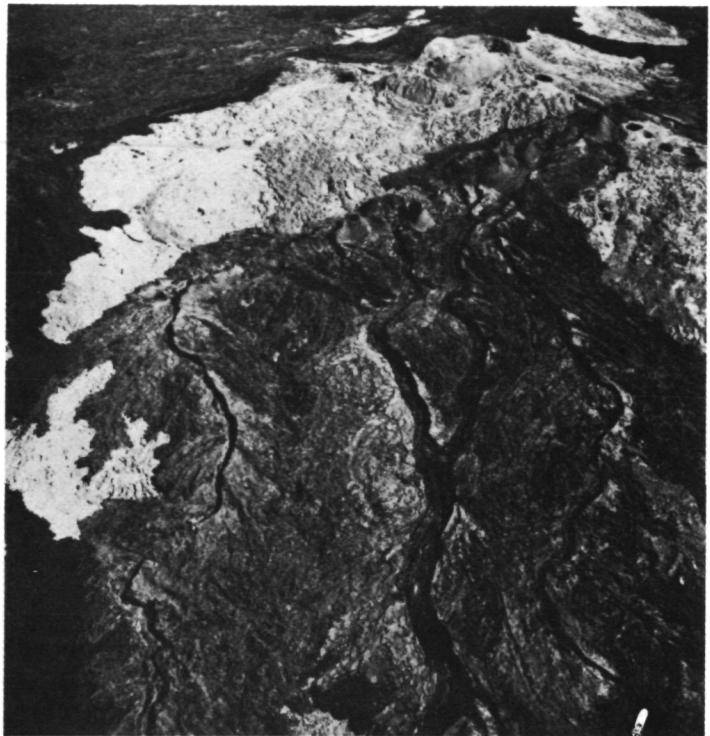
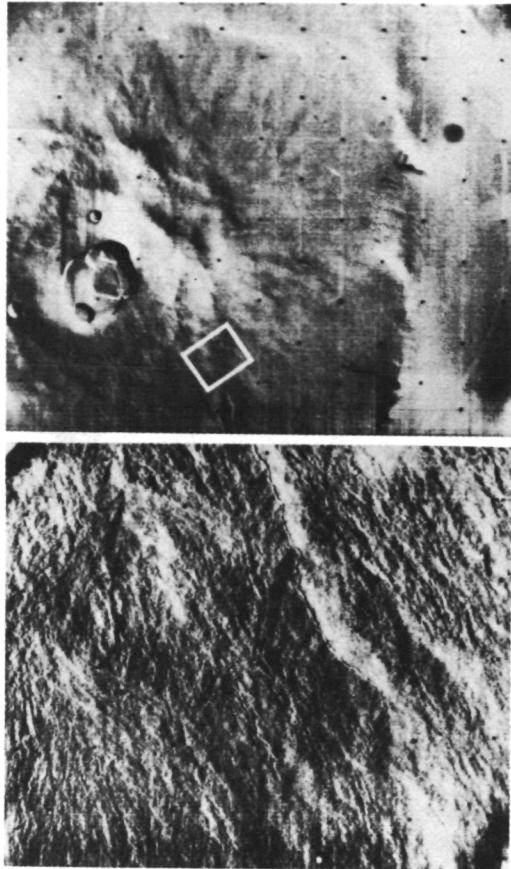


Figure 6.2. Comparative volcanology. Photograph in upper left shows part of Olympus Mons, a ~ 600 km-diameter martian shield volcano; lower left, high-resolution photograph of the flank of Olympus Mons showing numerous thin flows and a sinuous depression interpreted to be a lava channel. Compare the texture of the martian features with those observed on the flank of Mauna Loa shield volcano, Hawaii. The morphological similarities suggest that essentially the same process was involved in the formation of the martian shield as in the terrestrial analog, that is, construction by repeated eruptions of relatively fluid lavas.

Aouelloul Crater (Africa) are all raised-rim craters of similar geometry as the wind tunnel models. In all three cases, there occurs a distinctive pattern of wind erosion and sand deposition that is related to the general wind direction and that is essentially the same as reproduced in the laboratory.

Using the results from the wind-tunnel simulations and field studies of terrestrial analogs, theoretical modeling then permits an application to Mars where predictions can be made for the wind speeds needed to set particles in motion and thus the rates of wind erosion and deposition. Finally, taken in combination, the labora-

tory simulations, the field work, and the theoretical analysis permit the interpretation of spacecraft data toward the solution of one aspect of martian geology.

The example cited above serves to illustrate the interplay of several types of Earth-based investigations. The study is a multidisciplinary effort involving a geologist, a fluid dynamicist, and a planetary physicist. The complexities of planetary geology often require a multidisciplinary approach because seldom does any one investigator have sufficient background in all the required fields or in the combination appropriate for the solution of specific problems.



Figure 6.3. Crater vent and associated lava channel in the Snake River Plain, Idaho (left photograph; from Greeley and King 1974) compared with sinuous rilles near the Crater Prinz on the Moon. Although the lunar features are an order of magnitude larger, considerations of their morphology and the basaltic composition of the mare lavas indicate a similar origin.

In the sections that follow, specific details of the types of Earth-based studies involved in planetary geology are discussed.

FIELD STUDIES

D. U. Wise and G. E. McGill

Geologic exploration of other planetary bodies requires that a maximum amount of uniquely interpretable information be derived from available data. The interpretation of planetary history requires the identification of the

major and minor geologic agents that have modified planetary surfaces, the sequences of events on those surfaces, the ages of those events, and the rates at which geologic processes have operated. This requirement is the same as that faced by terrestrial geologists for several centuries in studies of the Earth, and one might assume that the only requirement for planetary geology is the simple application of terrestrial criteria to other bodies. However, terrestrial geologists have the ability to return to the field again and again to collect masses of data for their interpretations. Rarely have they had to concentrate on

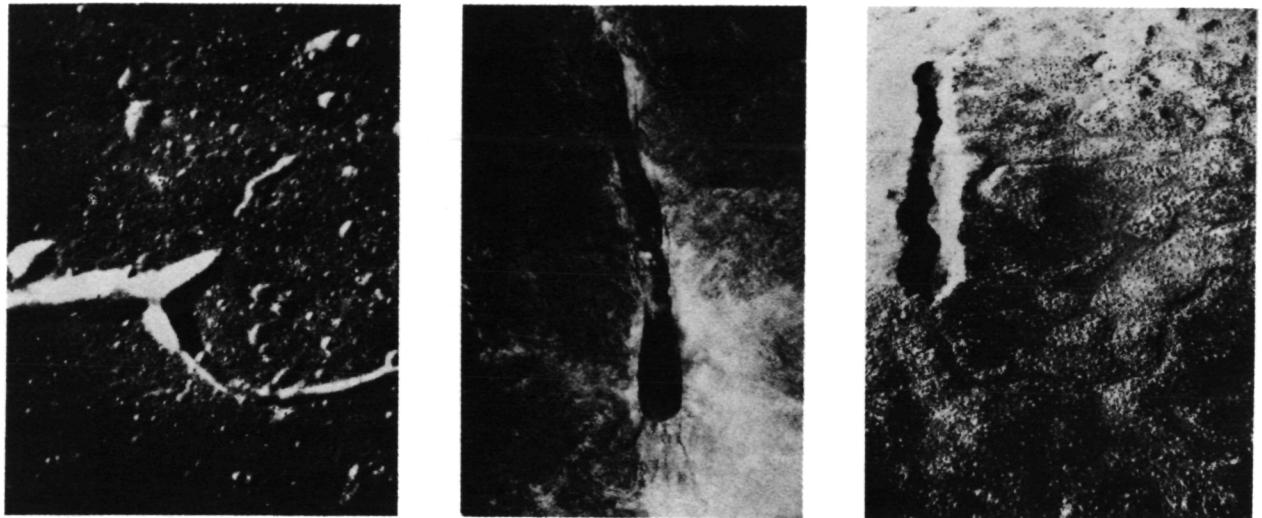


Figure 6.4. Elongate fissure-vent on the Moon (left), Hawaii (middle), and the Snake River Plain, Idaho (right).

Observations of the Hawaiian structure during eruptions over a period of several years show that such features evolve through a series of complex stages of multiple eruptions. Examination of the similar feature in Idaho suggests a similar history and enables interpretation of the lunar features. All three structures have lava tubes and channels associated with them.

finding characteristics that permit unique interpretations *solely* from remotely sensed data. Thus, the planetary geologist must re-examine many of the most obvious terrestrial landforms for criteria that lead to unique identifications, and for the details of landform evolution (Figs. 6.2–6.5).

Craters constitute one of the most common landforms found on planetary surfaces. The separation of impact craters from other circular depressions — mainly volcanic craters — is extremely important in planetary geology because many interpretations of planetary histories are based on impact crater frequency distributions. For example, heavily impact cratered surfaces are believed generally to be older than lightly cratered surfaces. Clearly, if nonimpact craters are mixed with the general crater population and cannot be identified, then an erroneously old age would be surmised.

Thus, the need arises for criteria that can be obtained remotely to identify various types of craters. Field studies of craters on Earth offer one means for deriving the desired criteria. It was through this approach that Shoemaker

studied Meteor Crater in Arizona and found ground features diagnostic of impact craters. Subsequent studies by many investigators of other impact craters, of volcanoes, and other types of craters have gradually built a body of knowledge that allows tentative crater identification. Unfortunately, the identifications still are not definitive, especially where planetary data are scant or where image resolution is poor, and field studies of terrestrial analogs must continue.

The fracture and stress history of planetary surfaces is recorded in some degree in the complex faults, linear volcanoes, and topographic lineaments that abound on the surfaces of all the planets, including Earth. On Earth, the origins of these fracture-related phenomena are poorly known and it is difficult to separate their origins from plate-tectonic phenomena. It is particularly difficult to separate the inherited effects of early fracture anisotropy of a planetary crust from fracture effects produced directly by younger stress fields. On planets with near-global rock exposures, longer preservation of topographic details, and no apparent continental drift, the answers to those problems may be more

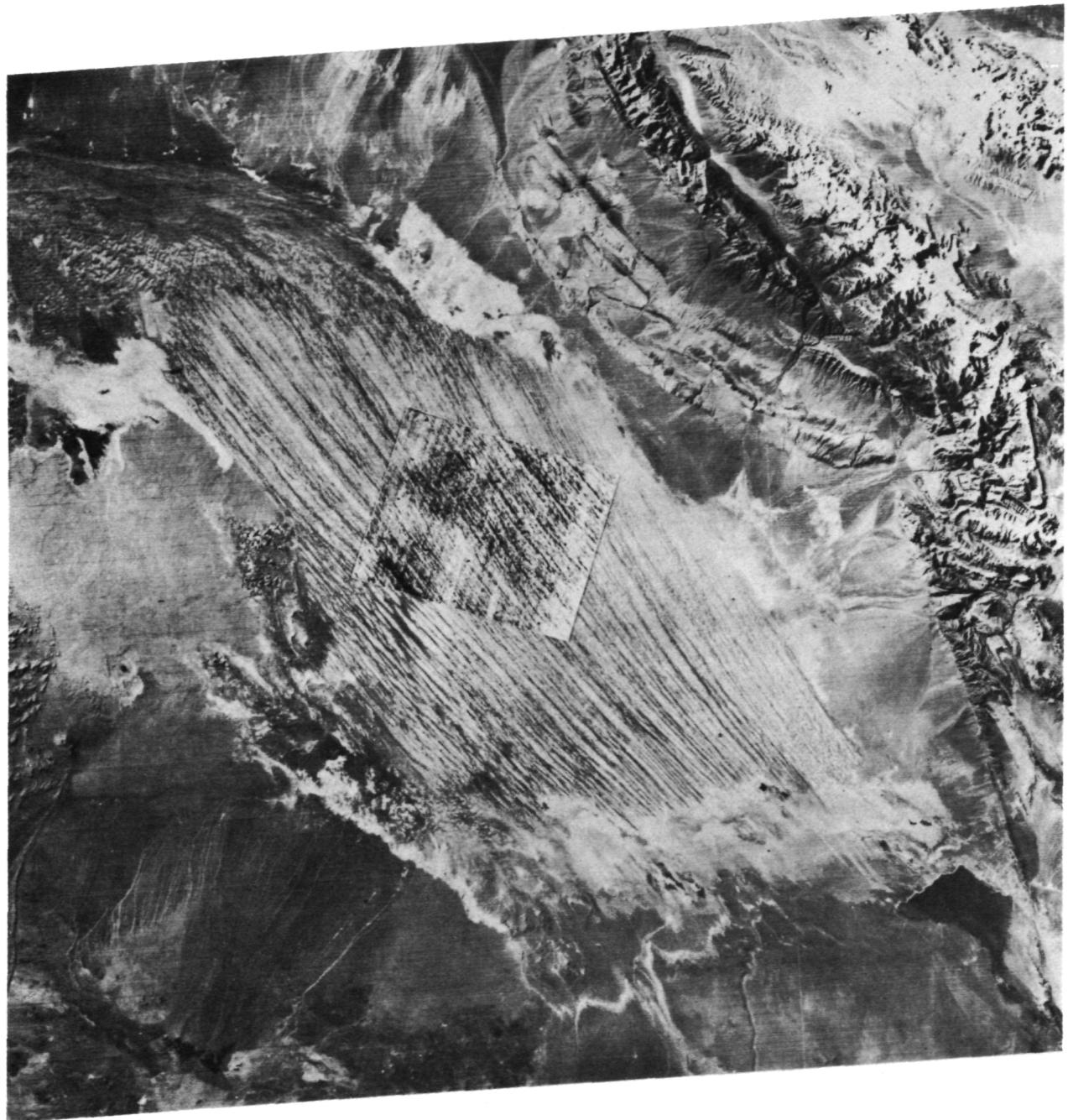


Figure 6.5. Composite of an ERTS picture and Mariner 9 frame to the same scale. The Mariner frame is DAS 6823253 from the southern Amazonis region of Mars and is about 50 km wide. The ERTS frame covers part of intensely wind eroded Lut Desert in Iran (from McCauley, in press).

obvious. At the very least, the fracture patterns of the other planets are forcing us to ask more sophisticated questions about terrestrial fractures and to seek the answers by field examina-

tion. In this area, a good possibility exists for making contributions to the understanding of terrestrial problems from comparative planetology.

Planetary geology, like many of the other planetary sciences, must have a laboratory in which to experiment and test various ideas. Physicists and astronomers often test many of their instruments and interpretations using terrestrial samples of rocks and minerals in laboratory experimentation. The planetary geologist has equally pressing requirements for laboratory testing of his ideas, particularly in that his interpretations commonly have less well-defined boundary conditions than those of the physicist or astronomer. This "laboratory" of Earth analogs is a key to the development of criteria for deriving the optimum amount of information from planetary surfaces.

Geologic interpretations of remotely sensed data provide a framework for planetary surface exploration. The study of terrestrial field examples, or Earth analogs, as laboratories for planetary geologic interpretations constitutes one of the least expensive but most important foundations of the planetary exploration program.

LABORATORY EXPERIMENTATION

A. Howard

Laboratory experimentation is a necessary companion to the development of theoretical models and comparative field studies in planetology. The analysis of the geologic framework of other planets by comparison with terrestrial environments and processes is handicapped by the vastly different environments and the lower resolutions, limited scope, and indirect nature of most of the data from other planetary bodies. Laboratory testing of theories in planetology is particularly difficult because of the problem of obtaining the controlled measurements necessary to validate theories. In some cases, terrestrial field observations can substitute for direct observation of planetary processes but only if analog environments occur on Earth. Even where these occur, quantitative observations may be difficult or impossible because of the multiplicity of active processes. Problems of logistics or instrumentation include the rarity or low intensity of certain processes, the complexity of

natural environments, large physical scales, or long time scales of the processes involved.

Laboratory experimentation often overcomes these limitations, forming a necessary link between field or planetary observations and theory construction. Experiments offer the advantage of providing quantitative analysis under controlled conditions because single processes may be isolated and chains of cause and effect unravelled. Under special circumstances, experimentation may be part of a field research program if the processes involved can be isolated and readily observed; that is, if nature provides the controlled conditions necessary for experimentation. For example, McGetchin and Head (1973) made photographic observations of volcanic ejecta forming cinder cones, thus providing empirical data that were combined with predicted trajectories under different gravity attractions and atmospheric resistances to form cinder cones on Earth, the Moon, and Mars. In a quite different context, Greeley *et al.* (1974) dug a crater in an Iowa cornfield to observe the deposition pattern of windblown snow and compared it with aeolian deposits around wind-tunnel models, around large terrestrial impact craters, and ultimately around martian craters.

However, suitably controlled natural experiments are rare. Laboratory experimentation often can overcome this problem, however, by providing isolated, idealized, and controlled environments and processes, and by allowing observations to be made on a reduced physical and temporal scale. In addition to the testing of formulated theories, laboratory and model experiments can be used to demonstrate effects of complex processes.

Meteoroid cratering is a dominant process in the evolution of solid-surface bodies in the solar system, as evidenced by the cratered surfaces of the Moon, Mars, Mercury, and the satellites of Mars, as well as the scars of large ancient impact events on Earth. Earth-based radar images shows that parts of Venus also are cratered. For atmosphere-free bodies, such as the Moon and Mercury, impact appears to have been the major agent of erosion and metamorphism. A thorough knowledge of cratering mechanics at all scales

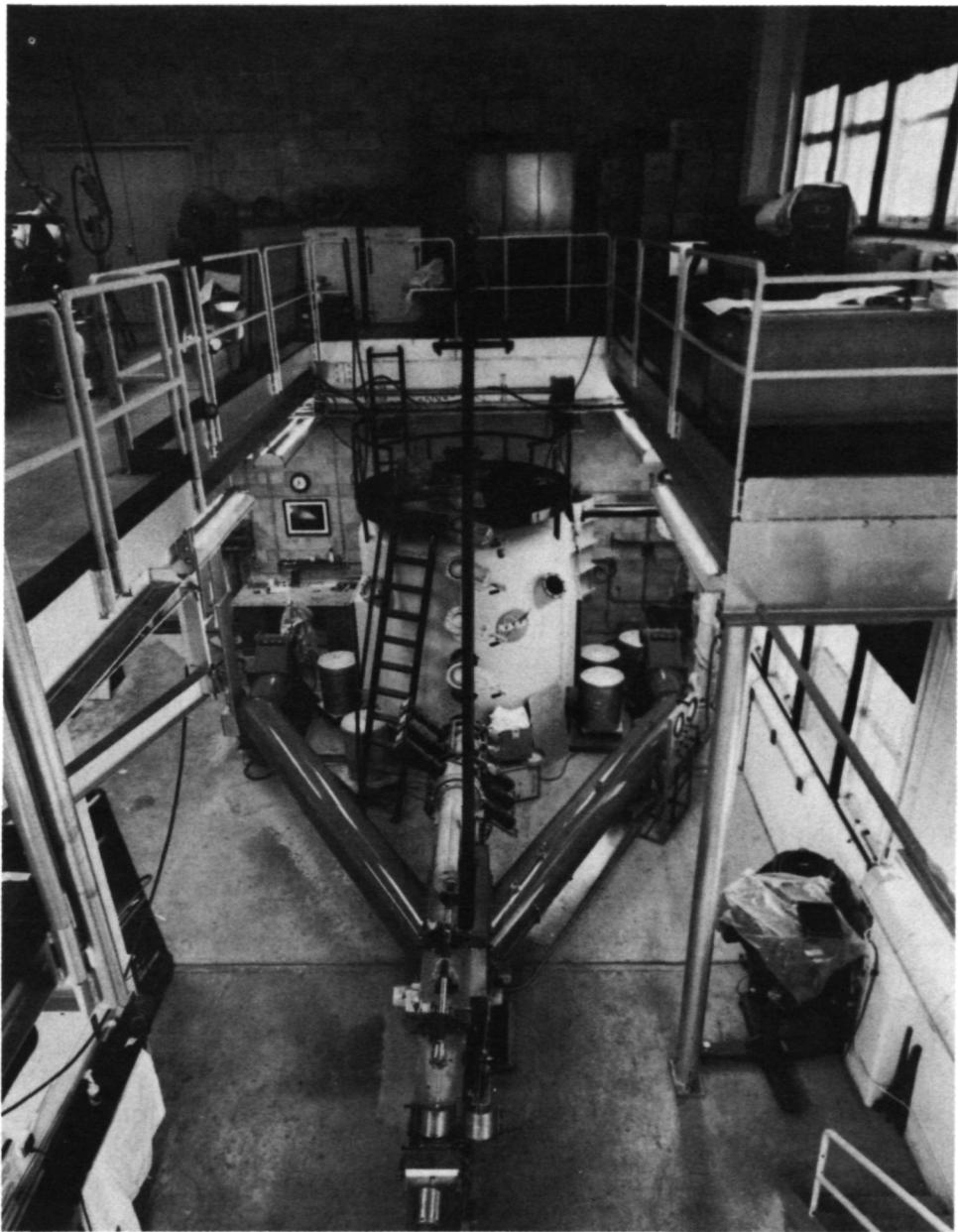


Figure 6.6. Vertical gun facility at Ames Research Center, used for simulating impact-cratering processes.
(Courtesy of D. Gault.)

and the influence of variables controlling crater formation and geometrics are absolute requirements for interpreting physical properties and structural details of the surfaces exposed to the meteoroid environment.

The impact-cratering process is controlled primarily by projectile properties (velocity, mass), target properties (strength, structure), and environmental properties (gravity, atmosphere).

These are parameters that can be duplicated, to a degree, in laboratory simulations. Because the conditions can be controlled and varied, each parameter can be assessed individually as to its role in the cratering processes. The cratering mechanics experiments by Gault (1974), using the Vertical Gas Gun at NASA-Ames Research Center (Fig. 6.6), classically illustrates the value of laboratory simulations. This facility is

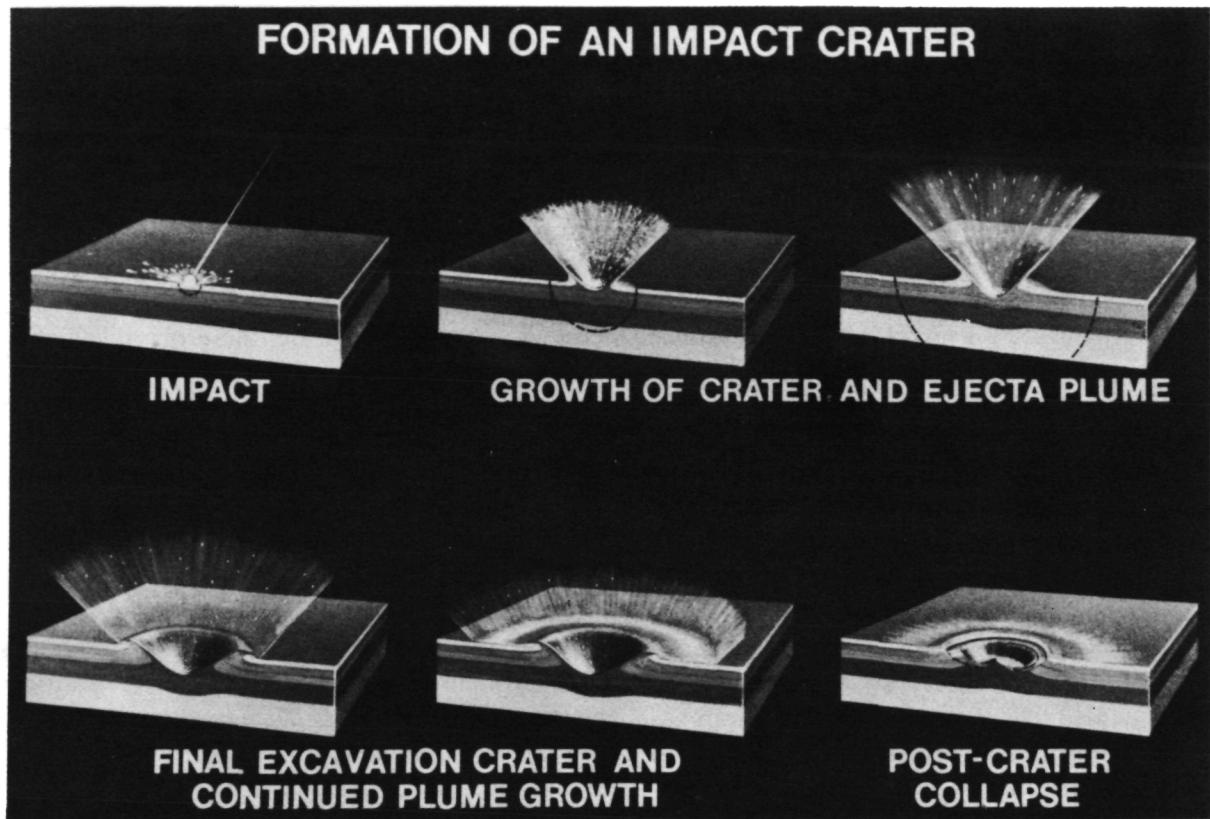


Figure 6.7 Diagrams illustrating the typical growth of an impact crater, compiled from analyses of hundreds of "stop-motion" photographs acquired during experiments using the facility shown in Figure 6.6. (Courtesy of D. Gault.)

capable of firing projectiles at speeds to 7.5 km/sec into targets under low atmospheric pressures suitable for simulating conditions on the Moon. At the time the projectile is fired, the target can be dropped at a controlled rate, reducing the effective gravitational acceleration to any desired value. The entire sequence of crater formation can be photographed by high-speed stereo motion picture and still cameras to provide detailed information on the various stages in crater formation (Fig. 6.7). Modeling techniques permit cautious extrapolation of various scaling laws from impact simulations to full-scale planetary structures.

The laboratory results are compared with terrestrial impact craters, both natural and man-made, and the "coordinated" results are applied to the interpretation of extraterrestrial craters. In some cases it is possible to separate exogenic

(impact) craters from those formed by endogenic processes, using a combination of morphological and distributional criteria. The impact craters then provide a baseline for interpreting the geomorphological evolution of a planetary surface as it is subjected to landforming processes, deposition, and erosion. Exogenic craters are formed almost at the same rate regardless of location on a planet, and if the source of cratering for all planets is the same, then the crater formation rate may have been varied the same. Statistical analysis of subsequent degraded crater morphologies then serve to elucidate the form and relative efficacy of various geomorphological processes on different parts of a planet's surface. The areal densities of craters of different sizes and degradational states further serve to date these processes relative to the cratering rate.

Laboratory experiments may proceed in two ways: (1) they may attempt to duplicate directly the environment and processes postulated to be present on a planetary body, such as studies of mineral stability and planetary interiors in alien surface environments, or (2) they may be performed in an environment, on a scale, or at an intensity different from the assumed planetary conditions and the results extrapolated by suitable scaling. Most laboratory studies of surface geologic processes, such as the aeolian transport of material on Mars, adopt the latter approach, owing to the difficulty in simulating the alien environment in the laboratory. For example, the reduced gravity and lower atmospheric pressure on Mars are difficult to simulate in the laboratory.

Whenever an exact duplication of a planetary environment is impossible, laboratory experiments should be conducted so that the results can be quantitatively extrapolated. Extrapolation is generally possible by appropriate analysis of a physical dimension of the materials and of the magnitudes of the processes in the laboratory and on the planetary body. Through the use of dimensionless ratios of physical quantities, laboratory experiments often can be performed so that, with proper combinations of velocities, pressures, temperatures, densities, physical dimensions, and other quantities, the experimental results can be appropriately extrapolated to the planetary environment. Often, however, certain physical effects, for example, the differences in atmospheric pressure between the laboratory and the planetary atmosphere, can be shown by experiment or theory to be unimportant to the process under study and therefore need not be scaled exactly. Where it is impossible to scale all relevant processes or environments, experiments still may provide qualitative or semiquantitative analogs of extraterrestrial processes, although more caution is necessary in the interpretation of results.

In addition to physical modeling of planetary processes, laboratory studies such as photometric and polarimetric analyses of samples and studies of ices and ice clathrates for the investigation of the surfaces of satellites in the outer

solar system are important aspects of laboratory experimentation in planetology.

Laboratory experimentation thus has an important role in studies of planetary geology and provide quantitative evaluation of geologic processes that supplement observations made by telescope, by space probes, and by terrestrial field studies. Laboratory studies allow a degree of control over processes and materials not otherwise possible and are therefore invaluable in testing theoretical models.

EXPERIMENTAL PETROLOGY

M. C. Gilbert

Introduction

Experimental petrology is the laboratory investigation of chemical systems thought appropriate in the understanding of petrogenesis. This means that chemical systems much simplified over natural ones can be studied as well as direct experimentation on representative natural samples. The goal of experimental petrology is generally elucidation of (1) the physiochemical conditions under which a rock formed, these being temperature, total pressure, fugacities of the various participating gaseous species, and activities of other components in the system, and (2) the specific reactions and reaction paths actually followed in rock formation. This goal is accomplished by duplicating, in heated pressurized vessels, the assemblage of minerals found significant in a particular natural setting. The value of experimental petrology thus lies precisely in the fact that it quantifies petrology.

Relevance of Experimental Petrology to Planetology

This topic is best approached by realizing first the contribution provided by experimental petrology to understanding the planet Earth. The following is a partial but representative list of these contributions:

1. Definition of the pressure-temperature

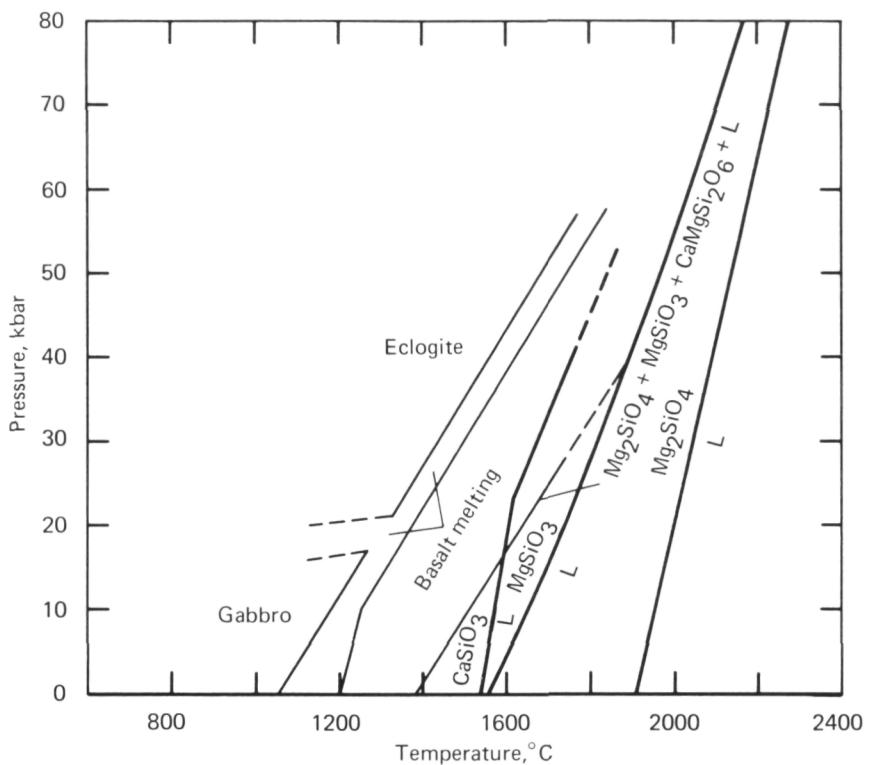


Figure 6.8. Selected silicate equilibria. Basalt melting curves as determined to 40 kbar under reducing conditions. Mg- and Ca-silicate melting curves represent univariant conditions. Data are compiled from various sources.

(P-T) fields of stability of common and diagnostic crustal mineral assemblages. This has allowed metamorphic petrology to advance beyond the concepts of isograds and facies to now include facies series (Miyashiro 1961; Hewitt and Gilbert 1975). The same approach is being extended to higher pressure assemblages thought characteristic of the upper mantle (MacGregor 1970).

2. Implicit in (1) is that the fossil geothermal gradient for a terrain can be estimated. Once this is known, tectonic models and thermal profiles can be discussed, all of which reflect more fundamental, deep-seated behavior in the upper mantle (Richardson 1970; Boyd 1973).

3. Mantle compositions have been inferred by partial melting experiments on various bulk compositions until liquids matching the range of natural lava types were obtained. This has also allowed determination of (a) the effect of different volatile constituents on melting behavior and (b) the fractionation of elements such as K, U, and Th between the participating solid phases

and liquid. The measured silicate melting curves are used to place constraints on the thermal state of the Earth's interior. Thus, the general processes of formation of differentiated crust, ocean, and atmosphere can now be quantitatively modeled (Boettcher 1975).

4. Relating seismically defined discontinuities to possible chemical and physical changes. All common rock-forming minerals of the Earth's crust undergo transformations to denser forms at high pressure. Only through laboratory experimentation can petrology and geophysics be merged into viable models of Earth structure and chemistry (Ringwood 1970).

5. Lava viscosities affect the form of volcanic edifices, and viscosity is a direct function of chemistry.

In principle, all of the points outlined are applicable to the terrestrial planets and to rocky satellites of the Jovian planets. The crucial question is, "To what extent can the laboratory studies of chemical systems be transferred to

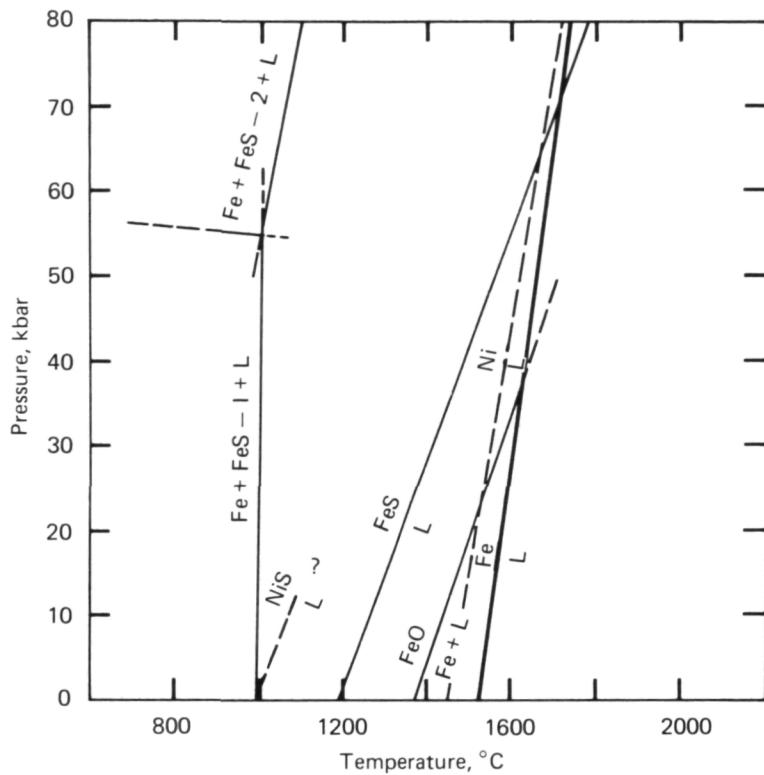


Figure 6.9. Selected equilibria involving metallic phases. Melting curves for the free metals, Fe and Ni, and their oxides and sulfides are shown. Data are from various sources.

other planets?" This question involves estimations of both the relevant bulk compositions and the appropriate physical conditions. It seems clear that much additional work will be necessary simply because evidence is abundant that bulk compositions are different from one planetary body to the next. Highly oxidizing conditions have rarely been investigated for Earth materials at high temperatures and pressures because such conditions were not applicable here. Yet such studies may be appropriate for other planets. In a similar vein, highly reducing conditions were not investigated before the advent of lunar studies as there was no clear reason to do so for the Earth. On the other hand, a substantial body of phase equilibrium data exists for many common minerals. Wherever these data have been determined for a wide range of physical conditions, they will probably be directly useful in planetology.

Experimental petrology has provided some interesting results for lunar interpretations

(Ringwood and Essene 1970; Green *et al.* 1971; Hays and Walker, in press, and others). Figures 6.8 and 6.9 depict some of the available melting equilibria that may prove applicable to the Moon and planets. The curves in Figure 6.8 are of materials representative of the stony mantles of the terrestrial planets. The curves in Figure 6.9 bear on possible core states and place limits on how early differentiation may have proceeded during planet heating. It is noteworthy that few good data are available at pressures over 50 kbar. Some of the data shown here have already been used by workers integrating geophysical constraints, thermal calculations, and petrology to give planetary models (e.g., Johnston *et al.* 1974).

For Mercury, different approaches to understanding the internal structure have led to conflicting ideas. Interpretation of the magnetic field based on Mariner 10 results points to a liquid core. Bulk geochemistry as inferred from some accretionary models points toward a

sulfur-free, Fe-rich core. The pressure at the core-mantle boundary is approximately 60 kbar. From Figure 6.9, melting temperatures would have to be above 1600° C. However, if sulfur were present, a liquid outer core could be maintained at \sim 1000° C. Thus, thermal and differentiation models are critically dependent on the bulk composition chosen. The choice of a mantle silicate geochemistry rich in Ca and Mg and low in Fe requires very high melting temperatures (Fig. 6.9).

For Mars, where some geochemical models imply high sulfur, a liquid core could be maintained at moderate temperatures. On the other hand, the basalt melting interval shown in Figure 6.8 may not be appropriate because of the possibility of higher oxidation state for the planet. If the higher oxidation state hypothesis were true, other consequences follow. Higgins and Gilbert (1973) have argued that Ni will then be released into the silicate system and be housed in the "ferromagnesian" minerals. Significant changes in melting temperatures, element fractionation, and phase stability are then possible. The higher volatile content postulated for Mars as compared with the Earth might render most of Figure 6.8 useless, in which case, other studies (e.g., Eggler 1974) are more pertinent.

METEORITES

T. Bunch

In the pre-Apollo era, much information about meteorites was gathered as an aid in the possible interpretation of the returned lunar samples. At that time, meteorites were the only extraterrestrial samples available for study, and most speculations about the origin of the solar system and the planets were based on information gained from meteorites and astronomical observations. Several investigators speculated that basaltic achondrites, because of their magmatic temperatures and Earth-like composition, came from the Moon. We have since learned that this is not the case, although the interest in meteorite research has not diminished, nor have we learned all that we can from meteorites

about the processes of the solar system (Fig. 6.10).

This section will briefly review the important role that meteorites have played in our quest for understanding the origin and nature of the solar system and what we can gain from the continued study of meteorites and their relatives, the asteroids. This discussion is based primarily on the geologic and chemical data from meteorites. Other important factors (isotope compositions, element distributions, rare gas contents, crystallization ages, geophysics, astronomical observations and measurements) are discussed elsewhere in this report.

Meteorites provide tangible samples of solar system material that has undergone compositional changes through differentiation and chemical fractionation. These changes probably occurred before and during emplacement of matter around the Sun, in the course of condensation of primitive solids from the early solar nebula gas cloud, and during changing magmatic and metamorphic processes in a parent body. Collisional shock events have also aided in altering the compositions of small planetary bodies.

Three types of meteorites are important for this discussion: (1) carbonaceous chondrites, which contain organic compounds, inclusions composed of refractory elements, and water; (2) ordinary chondrites, which contain chondrules formed by supercooling; and (3) basaltic achondrites, which have textures and mineralogy closely akin to lunar and terrestrial basalts and imply magmatic crystallization. The refractory element inclusions in carbonaceous chondrites may be an important clue to the condensation of the solar nebula. In discussing conditions of the solar nebula gas cloud, it is assumed that the composition of the cloud was heterogeneous, that is, the highest temperature refractory elements were concentrated closer to the proto Sun and decreased in abundance with increasing distance away from the Sun. Necessarily, the temperature gradient fell from highest temperature near the Sun to low temperatures near the outer portions of the primitive solar system. Because of the temperature gradient and the heterogeneous distribution of elements, it follows that all

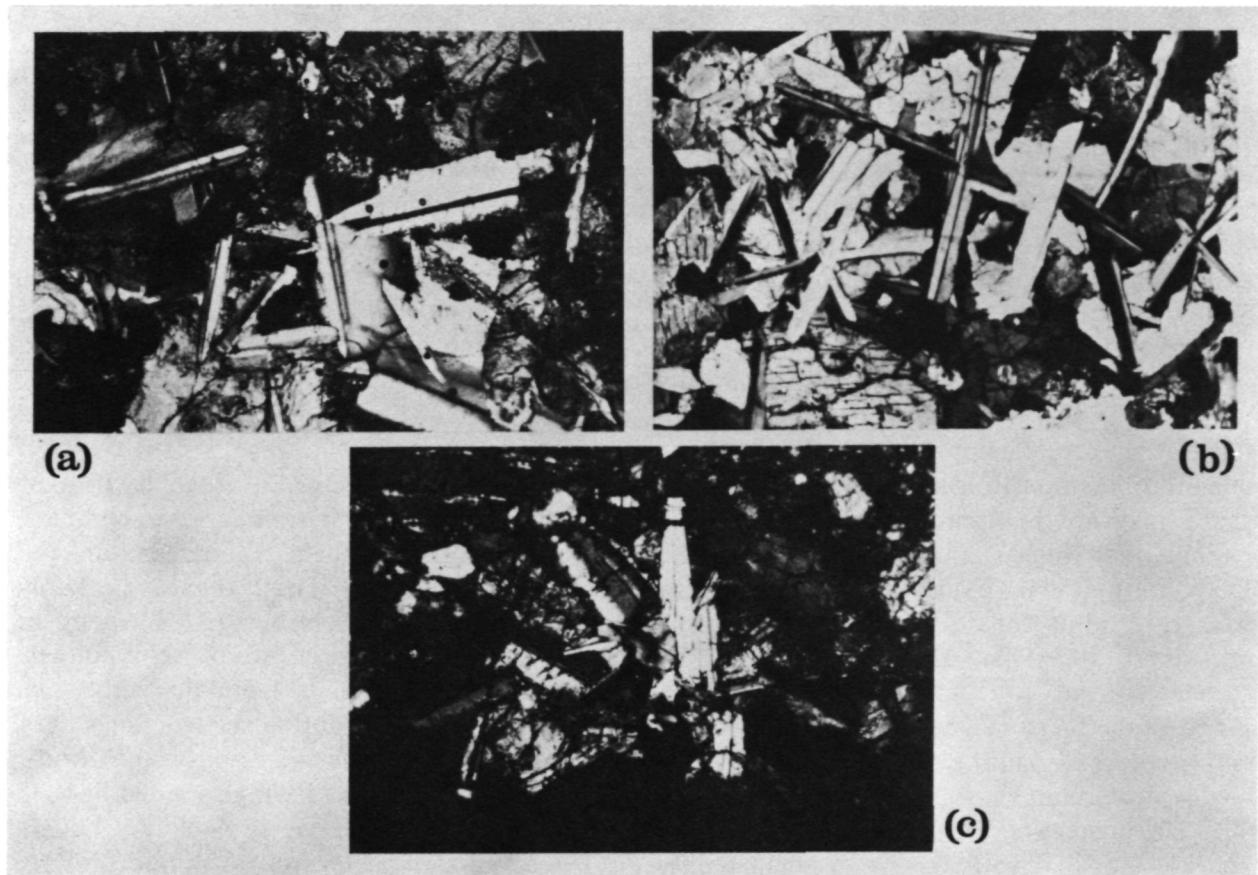


Figure 6.10. Comparative planetary evolution processes. Photomicrographs (crossed polarizers) of terrestrial (a), lunar (b), and meteoritic (c) basalts. All three are very similar in texture and mineral contents, although they are dissimilar in bulk composition and isotopic ratios. Terrestrial basalt age = 6×10^6 yr, lunar basalt age = 3.7×10^9 yr, and meteoritic basalt age = 4.4×10^9 yr. Width of field is 2 mm. Observational and experimental data demonstrate that many of the planetary bodies underwent similar geologic processes through time. Studies bearing on basalt generation from partial melting of primitive pre-existing materials can give insight on the composition of the parent bodies which, in turn, can be used to model the formational processes of the solar system.

elements were completely in the vapor state. Theoretical models and observational considerations imply that the individual planets change in composition from high concentrations of refractory and transitional elements near the Sun (terrestrial planets) to high concentrations of light, volatile elements in the colder environments farther from the Sun.

The presence of organic compounds in carbonaceous chondrites also indicates that low-temperature processes must have been involved

during certain phases of the history of these meteorites. The distribution of hydrocarbons, amino acids, and fatty acids suggests that these compounds have been formed by random, abiotic, chemical processes (Kvenvolden *et al.* 1970; Kvenvolden 1974). Stereochemical considerations of amino acids and fatty acids have now confirmed the idea that the processes were random and that the compounds do not result from extraterrestrial life. The physical conditions, time scales, and mechanisms of these low-

temperature processes remain to be determined, however, and this is an important direction for future research. Knowledge of the organic chemistry of meteorites will be critical in planetary exploration, especially in the interpretation of results from samples of the martian surface where organic compounds could be present from a number of sources including meteorites (see the section on organic chemistry in Ch. III).

A sequence of mineral condensation during the cooling of the nebular gas cloud has been calculated from thermodynamic data (Grossman and Larimer 1974) and is consistent with meteorite compositions, particularly refractory element inclusions in carbonaceous chondrites. In essence, the results suggest that the Ca-Al-rich inclusions are aggregates of highest temperature condensates. These condensates were followed, at lower temperatures, by crystallization of nickel-iron, forsterite (Mg_2SiO_4), and enstatite ($MgSiO_3$). With increasing oxidation state, iron substituted for magnesium in these minerals in ordinary chondrites. Thus, increasing the oxidation state and trends of decreasing abundance of siderophile elements appear to reflect formation at monotonically increasing distances from the Sun. The refractory condensates may have been transported by solar radiation pressure to other parts of the solar system before lower temperature condensation. This process may account for the admixture of both high- and low-temperature fractions in meteorites and for the inferred refractory element concentration of the Moon. Heterogeneous accumulation models may account for the apparent stratified nature of the Earth and possibly other terrestrial planets, where planetary interiors are enriched in refractory elements and iron, with the more volatile materials concentrated in the outer layers.

Ordinary chondrite compositions have long been regarded to represent at least an average meteoritic composition for the Earth. Chondritic abundances for major, minor, and trace elements have been used for standards of geochemical calibrations for the Earth and Moon. In addition, chondritic meteorites were formed very early in the history of the solar system. Precise crystallization ages of chondrites and

other meteorites from the $^{87}Rb-^{87}Sr$ method provide the best available measure of the age of the solar system.

Lunar studies strongly infer a post-accretional surface melting and differentiation stage that separated the Moon into an outer layer of refractory material (anorthositic rocks) and an underlying residuum that subsequently gave rise to massive basalt flows through partial melting. These events may not have been restricted to the Moon, but may have occurred on some or all of the terrestrial planets by rapid buildup of heat from concentration of extinct radionuclides or an exterior heating event during the T-Tauri solar wind (Sonnett *et al.* 1968) or the highly luminous Hayashi phase of the early Sun. The meteorite record shows that basaltic achondrites have magmatic textures and probably originated through partial melting of parent-body materials. Basaltic achondrite breccias (howardites) contain fragments of anorthositic composition, in addition to many kinds of basaltic fragments (Bunch 1975). Crystallization ages for two basalt fragments were determined by Papanastassiou *et al.* (1974) and they correspond to crystallization events of 3.6×10^9 and 3.9×10^9 years ago, events that are distinctively younger than ages for simple basaltic achondrite samples that average 4.4×10^9 years. These "young" magmatic events, together with textural studies, suggest that the parent body was geologically active for at least 1 aeon after accretion. The Nakhla meteorite is a highly differentiated rock that has an inferred crystallization age of 1.34×10^9 years (Papanastassiou and Wasserberg 1974). Nakhla also shows evidence of magmatic crystal settling and partial alteration by water (Bunch and Reid 1975; Reid and Bunch 1975). These conditions are similar to terrestrial processes and dissimilar to those of other extraterrestrial bodies, as we know them. From these cases, we can see that meteorite parent bodies have been geologically active for considerable periods of time, and we can gain tremendous insight into planetary processes by continued study of the many meteorite samples available.

Recent spectral reflectivity measurements of

asteroids (McCord and Gaffey 1974; Chapman 1975) show that many of the meteorite classes are represented in the surface materials. Most of the larger asteroids appear to have a Moon-like regolith. Meteorite breccias also appear to be similar to the lunar regolith and suggest that the parent bodies suffered severe surface modifications through impact events and mixing of materials. In addition, petrographic studies of these breccias (chondrite and achondrite breccias) indicate that each meteorite compositional and petrographic class probably originated as separate bodies because fragments of different meteorite classes are not found within other meteorite breccias; that is, H-group chondrites are not found in L-group chondrite breccias and ordinary chondrite fragments are not found in achondrite breccias as would be expected if these meteorites originated from the same parent body.

Summary

Meteorites have provided us and will continue to provide us firsthand knowledge of the geologic processes of the solar system. The questions raised by Adams *et al.* (1967 a,b) about the origin and evolution of the solar system can be dealt with, at least for the terrestrial group of planets, by applying our present knowledge of meteorites and the Moon, together with Earth-based observations and unmanned space probes. Continuing attempts to refine the answers to the Adams group questions will depend on our efforts to initiate more detailed and sophisticated studies of meteorites and the Moon and to develop space missions for further geologic study to substantiate the conclusions based on meteorite studies alone. The main limitation of the record from meteorites is that the source regions cannot yet be identified. Moreover, the relationship of meteorites among themselves in time and space also cannot be fully understood. Meteorites give us only fragments of information about the geologic evolution of the solar system on which we have based many models. We need to fill in the gaps by geologic missions to the planets, their satellites, and the asteroids.

GROUND-BASED TELESCOPIC OBSERVATIONS

E. Helin, D. Matson, and T. Johnson

Major advances to be made in the future concerning the geologic and geophysical aspects of the planets, satellites, and other solid bodies in the solar system will come from spacecraft observations; however, a great deal remains to be learned from ground-based telescopic observations. Both optical and radio telescopes have been used to lay the groundwork for most of what we know about the solar system today. The advantage of ground-based observations lies in the fact that they are vastly more economical than those conducted from spacecraft and can be extended over long periods of time. Furthermore, new ideas and new techniques can be exploited independently of the increasingly long lead times and planning that must go into the execution of a planetary spacecraft mission. Thus, any sound strategy to explore the planets should take full advantage of ground-based observations, both to add to the broader understanding of each solar system object and to aid in the selection, development, and refinement of the best achievable spacecraft experiments.

The selection of those ground-based telescopic observations that should be pursued most vigorously depends on the given states of the art of telescopic instruments and observing techniques. Emphasis should be placed on proven techniques that will provide a foundation of reliable knowledge and that will supplement, in critical ways, the data already obtained from spacecraft or the measurements expected to be obtained from spaceflight missions in the near future. Some of the most important telescopic observational programs needed for each type of solid body are listed in Table 6.1. The listing is necessarily incomplete because it is impossible to foresee all the observations that would be most rewarding.

TABLE 6.1. EARTH-BASED TELESCOPIC OBSERVATIONS OF THE SOLAR SYSTEM

| |
|--|
| Mercury |
| 1. Radar astronomy to obtain altimetry profiles |
| 2. Spectrophotometry in the wavelength region of 0.3 to 2.2 μm to study surface composition by the spectral reflectance method. |
| 3. Infrared and radio wavelength observations of thermal emission to investigate the temperature and physical properties of near-surface materials |
| Venus |
| 1. Radar astronomy for mapping of solid surface of the planet |
| Apollo/Amor Asteroids |
| 1. Systematic photographic search to discover new Apollo/Amor asteroids to improve estimates of the total population, types of orbits, and collision rates with the Moon and terrestrial planets |
| 2. Polarimetry and UBV photometry to determine albedo, size, period of rotation, and axial orientation |
| 3. Spectrophotometry to study surface composition |
| 4. Infrared spectrophotometry to obtain albedo, size and emissivity information |
| 5. Radar to study size, shape, and surface characteristics |
| Mars |
| 1. High-resolution imaging for synoptic weather observations |
| 2. Spectrophotometry in the 0.3-5 μm wavelength region to study composition of the surface, particularly distribution of bound water in the surface material |
| Martian Satellites |
| 1. Spectrophotometry to determine composition of surface material |
| 2. Infrared radiometry to determine thermal and physical properties of the near-surface materials |
| Main-Belt and Trojan Asteroids.— These objects probably represent remnants of original planetesimals. Their compositional differences are largely unstudied. These small bodies may not have been altered by endogenic processes and therefore may provide basic information on primordial matter and conditions extant in the early solar system. |
| 1. Photographic search for new asteroids of unusual or specially interesting families |
| 2. Lunar occultation observations to determine diameters |
| 3. Spectrophotometry (0.3-2.2 μm) to study surface composition |
| 4. Infrared wavelength radiometry and visual wavelength polarimetry to determine albedo and size |
| 5. Radar observations, which are now becoming feasible, to determine size, rotational periods, and positional data for the computation of precise ephemerides |
| Jupiter |
| Galilean Satellites |
| 1. High-resolution spectroscopy of Io and other satellites (e.g., measurement of sodium D line and its distribution and the search for other species) |
| 2. Spectrophotometry (0.3-5 μm) to study composition and distribution of water frost. |
| 3. Radar observations to obtain size, altimetry, and surface roughness |
| 4. Radio wavelength thermal emission observations to extend thermal data to depth and investigate model of thermal emission |
| 5. Theoretical studies to examine all data yielded by the various investigative techniques and produce models for thermal evolution and surface composition variation |
| 6. Laboratory studies of the effects of proton bombardment on minerals and rocks so the surface properties and the surface interaction with Jupiter's magnetosphere can be understood |
| Outer Satellites (Observational goals are similar to those for asteroids) |
| 1. Polarimetry |
| 2. Spectrophotometry |
| 3. Radiometry |

TABLE 6.1.— Concluded.

| | |
|-------------------------|---|
| Outer-Planet Satellites | <ol style="list-style-type: none"> 1. Spectrophotometry to study surface composition in the 0.3-2.2μm wavelength region 2. Infrared photometry and radio emission to derive albedos and sizes and surface characteristics 3. Lunar occultation observations to determine diameters and possibly limb darkening |
| Rings of Saturn | <ol style="list-style-type: none"> 1. Spectrophotometry (0.3-5 μm) to examine distribution of color and composition over rings 2. Radio and radar observations to investigate particle-size distribution and composition 3. Infrared radiometry to determine albedo, radii, and temperature distribution |
| Pluto | Spectrophotometry (0.3-2 μ m) to obtain surface compositional information |
| Comet Nuclei | An integrated effort employing all observational techniques to investigate comets that no longer show activity, and comets at large heliocentric distances should be undertaken to study the physical and chemical properties of cometary nuclei (Comet Arend-Rigaux is one possibility in the next few years). |
| Orbit-Evolution.— | These studies provide the time-history relationship between planetary satellites and asteroidal objects. The problems relate to: <ol style="list-style-type: none"> 1. Origin of the bodies providing the early heavy bombardment in the solar system as recorded by the surfaces of the Moon, Mars, and Mercury 2. Origin of meteorites and their parent bodies 3. Evolution of satellite systems 4. Origin of comets 5. The unusual orientation of Uranus' rotation axis |

TERRESTRIAL RADAR OBSERVATIONS

G. Schaber

From a geological viewpoint, the most important aspect of radar is that it provides a means to map the topography of a planet whose surface is obscured by an atmosphere. This is particularly important for Venus (Fig. 6.11), which is very similar in radius and density to the Earth and so might reasonably be expected to have followed a similar evolutionary path. We know, however, that the atmosphere of Venus is drastically different from the Earth's, and it is essential to know whether the solid parts of the planets exhibit comparable contrasts. Comparison of the topography of Venus with that of the Earth is an essential first step in evaluating the nature and causes of difference between the two planets. The presence on Venus of orogenic belts, rift zones, and sedimentary basins similar to those on Earth will have particular geologic significance. The necessary images could be obtained from either terrestrial measurements or orbital measurements (see chapter entitled Stratigraphy and Structural Geology). It is impor-

tant, however, to obtain both types of image data. (See section on Radar Images, Ch. 2.)

Delay-Doppler radar images, with spatial resolution comparable to telescopic optical photography (approximately 2 km) have already been obtained of the Moon. With improved receivers and data-handling techniques, comparable resolutions are achievable for a small portion of Venus from terrestrial observations. Images for some areas have been obtained with existing radar interferometer techniques down to 10–15-km resolution (Rumsey *et al.* 1974). The large Arecibo radar dish in Puerto Rico, recently upgraded with a new dish netting, shorter wavelength transmitter (12.5 cm), and refined electronics, is expected to obtain greatly improved 2–10-km resolution images for parts of Venus in the future. Resolution is limited primarily by the signal/noise ratio, and it is expected that this ratio can be significantly improved for all planetary radar dishes in the near future.

Support of terrestrial radar observation of Venus with particular emphasis on improving spatial resolution is highly recommended. This has higher priority than any other measurement

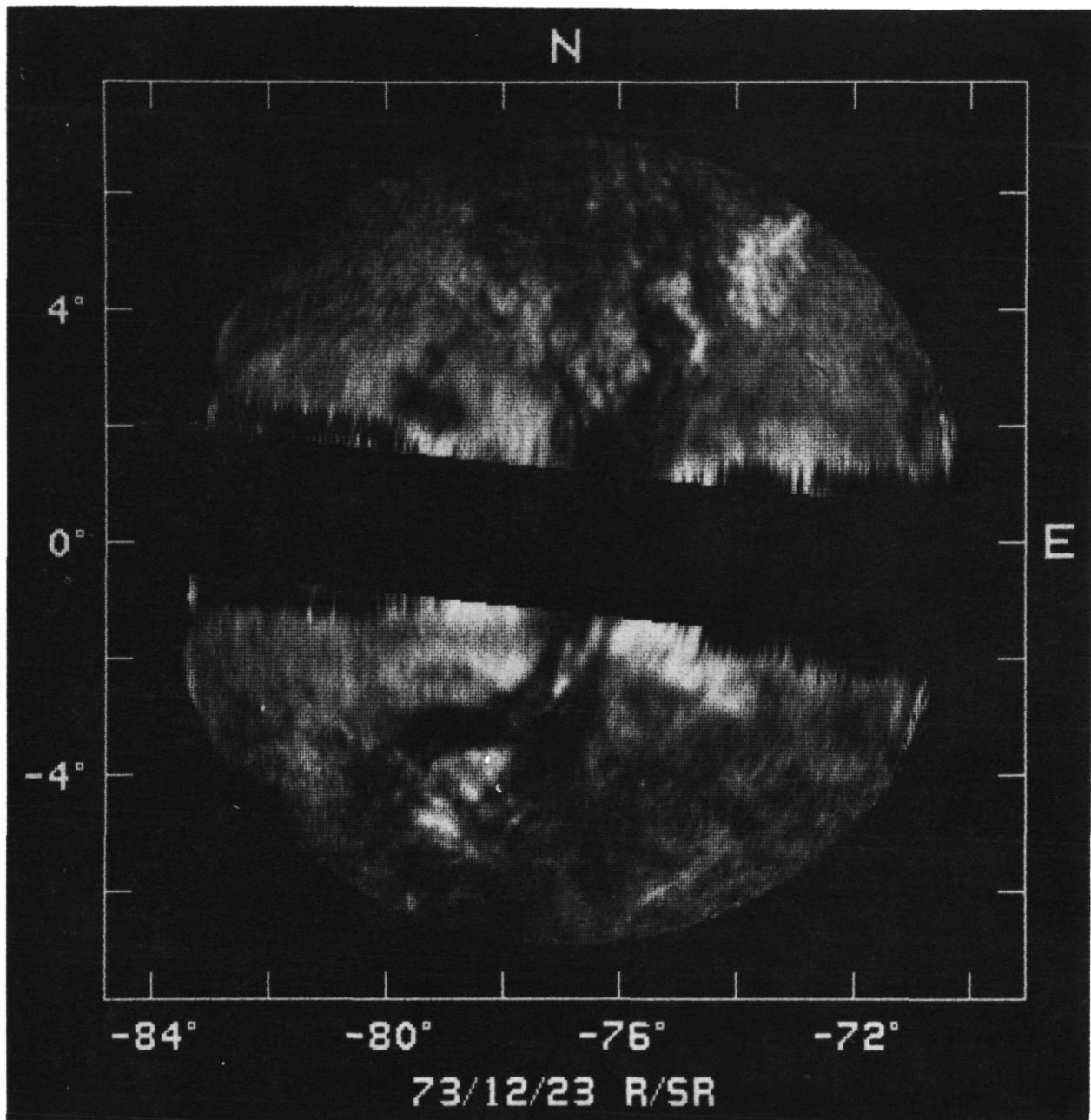


Figure 6.11. Radar image of the surface of Venus acquired at the Goldstone Tracking Station. The image shows an area approximately 1500 km in diameter. A large, trough-like feature (~1500 km long by 150 km wide by 2-4 km deep) is seen near the center of the frame. The main linear branch of the trough extends some 650 km trending NNE-SSW and is centered near the equator and -76° east longitude. North of this linear branch the single-channel form deteriorates into several north-trending, roughly parallel troughs, the largest of which has scalloped walls, suggesting a crater-chain-like appearance. To the south, the trough divides into two branches, each about 100 km wide. A ridge or septum appears to run the length of the linear portion of the trough system. Surrounding the troughs are essentially featureless plains. By analogy with the East African/Ethiopian rift system on Earth and the Valles Marineris canyon system on Mars, the Venusian feature could be interpreted to be the result of extensional tectonics. Whether the trough is associated with spreading center phenomena or simply an extension of the crust cannot be determined at this time. The possibility of studying plate tectonic phenomena in a nonaqueous and possibly pristine form makes Venus an exciting candidate for future exploration (radar image courtesy of R. Goldstein, Jet Propulsion Lab., caption by M. Malin, Jet Propulsion Lab.).

of Venus. A necessary supporting effort is to improve knowledge of the rotation and orbital constants of the planets, particularly of Venus.

THEORETICAL STUDIES

Because it is unlikely that enough definitive data will ever be in hand to fully understand the complex geology and history of solar system objects, we must be able to extrapolate from the known to the unknown. Such extrapolations are done through theoretical studies, that is, numerical modeling, simulations, and so forth. Given certain values for various parameters obtained by measurement or from "best case" estimates as a starting point, theoretical modeling produces possible solutions. Through numerical simulations, individual parameters can be varied to determine cause and effect relations. As more observations are made and measurements obtained, constraints are placed on the models and the number of possible solutions is reduced. As we have discussed in this chapter, measurements for these constraints come from many sources — spacecraft, terrestrial field studies, laboratory simulations, rock samples, and so forth.

For example, for many years, planetologists have derived models for planetary interiors using known values such as planetary diameter and

mass as starting points. When compositional data for the Moon became available, constraints were placed on lunar interior models. Additional data from seismometers, magnetometers, and heat-flow experiments placed further constraints on theoretical models and the number of possibilities was reduced. Similar exercises have been performed for Mars and Mercury and although the type and amount of data are not nearly as extensive as for the Moon, the knowledge gained from the Moon itself places some constraints on models of other terrestrial objects.

Another topic of high interest among planetologists is the manner in which lavas of essentially the same composition behave under different planetary conditions. Using measurements obtained from studies of both active and cooled lavas on Earth, several investigators, through theoretical modeling and substitution of appropriate parametric values, have applied the results to interpret lava flow characteristics on the Moon (e.g., Schaber 1973) and the formation of sinuous rilles (Hulme 1973; Carr 1974).

It is not the intent to discuss here all the ramifications of theoretical studies. In practice, nearly all aspects of planetology require theoretical solutions to some degree. Geological measurements provide the necessary constraints that ultimately limit the solutions derived from theoretical studies.

REFERENCES CITED

Adams, J. B., Conel, J. E., Dunne, J. A., Fanale, F., Holstrom, G. B., Loomis, A. A., 1967a. A strategy for scientific exploration of the terrestrial planets. *Rev. Geophys.* 7:623–661.

Adams, J. B., Conel, J. E., Dunne, J. A., Fanale, F., Holstrom, G. B., Loomis, A. A., 1967b. Scientific strategy, objectives and experiments for Mariner missions from 1970 through 1973. Calif. Inst. Tech. Jet Prop. Lab. Doc. 900-94 (internal document).

Boettcher, A. L., 1975. Experimental igneous petrology. *Rev. Geophys. Space Phys.* 13(3):75–79, 117–120.

Boyd, F. R., 1973. A pyroxene geotherm. *Geochim. Cosmochim. Acta* 37:2533–2546.

Bunch, T. E., 1975. Basaltic achondrite polymict breccias: an interpretation based on lunar surface breccia criteria: Abstracts, *Lunar Science VI*, The Lunar Science Institute, Houston, Texas, pp. 106–108.

Bunch, T. E., and Reid, A. M., 1975. The Nakhrites — Part I: Petrography and mineral chemistry. *Meteoritics* 10:303–315.

Carr, M. H., 1974. The role of lava erosion in the formation of lunar rilles and martian channels. *Icarus* 22:1–23.

Chapman, C. R., 1975. The nature of asteroids. *Scientific Am.* 232:24-33.

Eggler, D. H., 1974. Application of a portion of the system $\text{CaAl}_2\text{Si}_2\text{O}_8$ - $\text{NaAlSi}_3\text{O}_8$ - SiO_2 - MgO - Fe-O_2 - H_2O - CO_2 to genesis of the calc-alkaline suite. *Am. J. Sci.* 274:297-315.

Gault, D., 1974. Impact cratering in R. Greeley and P. Schultz, eds., *Primer in Lunar Geology*, NASA TM X-62,359, pp. 137-189.

Greeley, R., Iversen, J. D., Pollack, J. B., Udovich, N., and White, B., 1974. Wind-tunnel simulations of light and dark streaks on Mars. *Science* 183:847-849.

Greeley, R., Iversen, J. D., Pollack, J. B., Udovich, N., and White, B., 1974. Wind-tunnel studies of martian aeolian processes. *Proc. Roy. Soc. London A.* 341:331-360.

Greeley, R., and King, J. S., 1975. Geologic field guide to the Quaternary volcanics of the south-central Snake River Plain, Idaho. Idaho Bur. Mines and Geology, Pamphlet 160.

Green, D. H., Ringwood, A. E., Ware, N. G., Hibberson, W. O., Major, A., and Kiss, E., 1971. Experimental petrology and petrogenesis of Apollo 12 basalts. *Proc. Second Lunar Science Conf.*, 1:601-616. *Geochim. Cosmochim. Acta*, Suppl. 1.

Grossman, L., and Larimer, J. W., 1974. Early chemical history of the solar system. *Rev. Geophys. Space Phys.* 12:71-101.

Hays, J. F., and Walker, D., in press. Lunar igneous rocks and the nature of the lunar interior. *Proc. Soviet-American Conf. on Cosmochemistry of the Moon and Planets*, Pomeroy, J. H., ed., NASA, 1975.

Hewitt, D. A., and Gilbert, M. C., 1975. Experimental metamorphic petrology. *Rev. Geophys. Space Phys.* 13(3):79-81, 120-128.

Higgins, B. B., and Gilbert, M. C., 1973. High pressure stability of nickel diopside. *Am. J. Sci.* 273-A:511-521.

Hulme, G., 1973. Turbulent lava flow and the formation of lunar sinuous rilles. *Mod. Geol.* 4:107-117.

Johnston, D. H., McGetchin, T. R., and Toksoz, M. N., 1974. The thermal state and internal structure of Mars. *J. Geophys. Res.* 79:3959-3971.

Kvenvolden, K. A., 1974. Amino and fatty acids in carbonaceous meteorites. In Dose, K., Fox, S. W., Deborin, G. A., and Pavlovskaya, T. E., eds., *The Origins of Life and Evolutionary Biochemistry*, New York: Plenum Publishing Corp.

Kvenvolden, K. A., Lawless, J., Pering, K., Peterson, E., Flores, J., Ponnampерuma, C., Keplan, I. R., and Moore, C., 1970. Evidence for extraterrestrial amino acids and hydrocarbons in the Murchison meteorite. *Nature* 228:923-926.

MacGregor, I. D., 1970. The effect of CaO , Cr_2O_3 , Fe_2O_3 , and Al_2O_3 on the stability of spinel and garnet peridotites. *Phys. Earth Planet. Interiors* 3:372-377.

McCord, T. B., and Gaffey, M. J., 1974. Asteroids: surface composition from reflection spectroscopy. *Science* 186:352-355.

McCauley, J. F., in press. The surficial geology of Mars. In *Proc. Colloquium on Planetary Physics and Geology*, Acad. Nazionale Sincei, Rome, 1974.

McGetchin, T. R., and Head, J. W., 1973. Lunar cinder cones. *Science* 180:68-71.

Miyashiro, A., 1961. Evolution of metamorphic belts. *J. Petrol.* 2:277-311.

Papanastassiou, D. A., Rajan, R. S., Hunke, J. C., and Wasserberg, G. J., 1974. Rb-Sr ages and lunar analogs in a basaltic achondrite; implication for early solar system chronologies. Abstracts. *Lunar Science VI*, The Lunar Science Institute, Houston, Texas, pp. 583-585.

Papanastassiou, D. A., and Wasserberg, G. J., 1974. Evidence for late formation and young metamorphism in the achondrite Nakhla. *Geophys. Res. Letters* 1:23-26.

Reid, A. M., and Bunch, T. E., 1975. The Nakhrites - Part II: Where, when and how. *Meteoritics* 10:317-323.

Richardson, S. W., 1970. The relation between a petrogenetic grid, facies series and the geo-thermal gradient in metamorphism. *Fortschr. Mineral.* 47:65–76.

Ringwood, A. E., 1970. Phase transformations and the constitution of the mantle: *Phys. Earth Planet. Interiors* 3:109–155.

Ringwood, A. E., and Essene, E., 1970. Petrogenesis of Apollo 11 basalts, internal constitution and origin of the moon. Proc. First Lunar Science Conf., 1:769–800. *Geochim. Cosmochim. Acta*, Suppl. 1.

Rumsey, H. C., Morris, G. A., Green, R. R., and Goldstein, R. M., 1974. A radar brightness and altitude image of a portion of Venus. *Icarus* 23:1–7.

Schaber, G. G., 1973. Lava flows in Mare Imbrium: Geologic evaluation from Apollo orbital photography. Proc. Fourth Lunar Sci. Conf., *Geochim. Cosmochim. Acta*, Suppl. 4 1:73–92.

Sonnett, C. P., Colburn, D. S., and Schwartz, K., 1968. Electrical heating of meteorite parent bodies and planets by dynamo induction from a premain sequence T-Tauri “solar wind”. *Nature* 219:924–926.

APPENDIX I. CONTRIBUTING AUTHORS

| Author | Address |
|------------------|---|
| Adams, J. B. | Department of Earth and Planetary Sciences, 24-422, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139 |
| Anderson, D. M. | Earth Sciences Branch, U.S. Army Cold Regions Research and Engineering Laboratory, P.O. Box 282, Hanover, New Hampshire 03755 |
| Arthur, D. W. G. | Center of Astrogeology, U. S. Geological Survey, 601 East Cedar Ave., Flagstaff, Arizona 86001 |
| Batson, R. M. | Center of Astrogeology, U. S. Geological Survey, 601 East Cedar Ave., Flagstaff, Arizona 86001 |
| Bunch, T. E. | Ames Research Center, NASA, Moffett Field, California 94035 |
| Carr, M. H. | Branch of Astrogeologic Studies, U. S. Geological Survey, 345 Middlefield Road, Menlo Park, California 94025 |
| Cassen, P. M. | Ames Research Center, NASA, Moffett Field, California 94035 |
| Chapman, C. R. | Planetary Science Institute, 252 West Ina Road, Suite D, Tucson, Arizona 85704 |
| Cuzzi, J. | Ames Research Center, NASA, Moffett Field, California 94035 |
| Dwornik, S. E. | NASA Headquarters, Washington, D. C. 20546 |
| England, A. W. | U. S. Geological Survey, Federal Center, Denver, Colorado 80225 |
| Fanale, F. P. | Jet Propulsion Laboratory, 4800 Oak Grove Drive, Pasadena, California 91103 |
| Gilbert, M. C. | Department of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, Virginia 24061 |
| Greeley, R. | University of Santa Clara, <i>at</i> NASA-Ames Research Center, Mail Stop 245-5, Moffett Field, California 94035 |
| Guest, J. E. | University of London Observatory, Mill Hill Park, London NW7, 2QS, England |

| | |
|-------------------|---|
| Helin, E. | Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91109 |
| Howard, A. D. | Department of Environmental Sciences, University of Virginia, Charlottesville, Virginia 22903 |
| Howard, H. T. | Stanford University, Radio Science Laboratory, Stanford, California 94305 |
| Huguenin, R. L. | Remote Sensing Laboratory, Department of Earth and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139 |
| Johnson, T. | Jet Propulsion Laboratory, 4800 Oak Grove Drive, Pasadena, California 91103 |
| Kieffer, H. H. | Department of Planetary & Space Science, University of California, Los Angeles, California 90024 |
| Kvenvolden, K. A. | U. S. Geological Survey, 345 Middlefield Road, Menlo Park, California 94025 |
| Langseth, M. G. | Lamont-Doherty Geological Observatory, Columbia University, Palisades, New York 10964 |
| Matson, D. | Jet Propulsion Laboratory, 4800 Oak Grove Drive, Pasadena, California 91103 |
| McCord, T. B. | Department of Earth and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139 |
| McGill, G. E. | Department of Geology, University of Massachusetts, Amherst, Massachusetts 01002 |
| Metzger, A. E. | Jet Propulsion Laboratory, 4800 Oak Grove Drive, Pasadena, California 91103 |
| Morris, E. | Center of Astrogeology, U. S. Geological Survey, 601 East Cedar Ave., Flagstaff, Arizona 86001 |
| Murray, B. C. | Division of Geological and Planetary Sciences, California Institute of Technology, 1201 East California Boulevard, Pasadena, California 91109 |
| Saunders, S. | Jet Propulsion Laboratory, 4800 Oak Grove Drive, Pasadena, California 91103 |
| Schaber, G. | Center of Astrogeology, U. S. Geological Survey, 601 Cedar Ave., Flagstaff, Arizona 86001 |

Steggert, M. A. Space Research Coordination Center, University of Pittsburgh, Pittsburgh, Pennsylvania 15260

Toulmin, P., III Geologic Division, U. S. Geological Survey, Washington, D. C. 20242

Tyler, G. L. Center for Radio Astronomy, Stanford University, Stanford, California 94305

Wilhelms, D. E. Branch of Astrogeologic Studies, U. S. Geological Survey, 345 Middlefield Road, Menlo Park, California 94025

Wise, D. U. University of Massachusetts, Department of Geology, Amherst, Massachusetts 01002

Zartman, R. E. U. S. Geological Survey, Federal Center, Denver, Colorado 80225

NATIONAL AERONAUTICS AND SPACE ADMINISTRATION
WASHINGTON, D.C. 20546

OFFICIAL BUSINESS
PENALTY FOR PRIVATE USE \$300

SPECIAL FOURTH-CLASS RATE
BOOK

POSTAGE AND FEES PAID
NATIONAL AERONAUTICS AND
SPACE ADMINISTRATION
451



POSTMASTER : If Undeliverable (Section 158
Postal Manual) Do Not Return

"The aeronautical and space activities of the United States shall be conducted so as to contribute . . . to the expansion of human knowledge of phenomena in the atmosphere and space. The Administration shall provide for the widest practicable and appropriate dissemination of information concerning its activities and the results thereof."

—NATIONAL AERONAUTICS AND SPACE ACT OF 1958

NASA SCIENTIFIC AND TECHNICAL PUBLICATIONS

TECHNICAL REPORTS: Scientific and technical information considered important, complete, and a lasting contribution to existing knowledge.

TECHNICAL NOTES: Information less broad in scope but nevertheless of importance as a contribution to existing knowledge.

TECHNICAL MEMORANDUMS: Information receiving limited distribution because of preliminary data, security classification, or other reasons. Also includes conference proceedings with either limited or unlimited distribution.

CONTRACTOR REPORTS: Scientific and technical information generated under a NASA contract or grant and considered an important contribution to existing knowledge.

TECHNICAL TRANSLATIONS: Information published in a foreign language considered to merit NASA distribution in English.

SPECIAL PUBLICATIONS: Information derived from or of value to NASA activities. Publications include final reports of major projects, monographs, data compilations, handbooks, sourcebooks, and special bibliographies.

TECHNOLOGY UTILIZATION PUBLICATIONS: Information on technology used by NASA that may be of particular interest in commercial and other non-aerospace applications. Publications include Tech Briefs, Technology Utilization Reports and Technology Surveys.

Details on the availability of these publications may be obtained from:

SCIENTIFIC AND TECHNICAL INFORMATION OFFICE

NATIONAL AERONAUTICS AND SPACE ADMINISTRATION

Washington, D.C. 20546